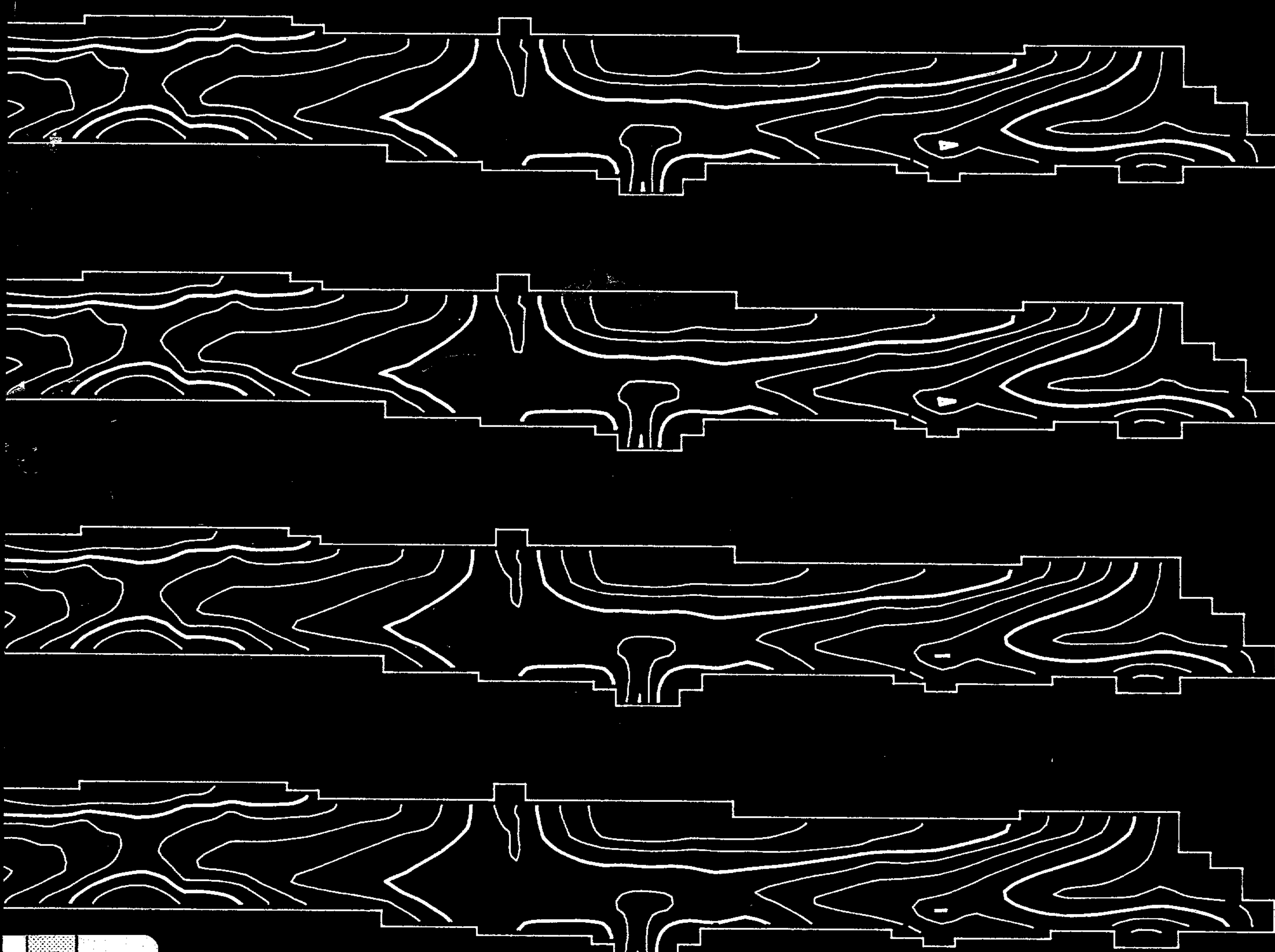
Water-Resources Investigations Report 90-4008

HYDROGEOLOGY AND GEOCHEMISTRY IN BEAR CREEK AND UNION VALLEYS NEAR OAK RIDGE, TENNESSEE



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.S. GEOLOGICAL SURVEY

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.S. DEPARTMENT OF ENERGY

Cover: Layers 1-4, from bottom to top, of the simulated ground-water-flow model in Bear Creek and Union Valleys, near Oak Ridge, Tennessee.

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By Zelda Chapman Bailey and Roger W. Lee

U.S. GEOLOGICAL SURVEY

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Nashville, Tennessee

1991

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CONVERSION FACTORS

Multiply inch-pound unit	By	To obtain metric unit
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
acre	0.4047	hectare
foot per day (ft/d)	0.3048	meter per day
inch per year (in/yr)	2.54	centimeter per year
foot squared per day (ft^2/d)	0.0929	meter squared per day
cubic foot per day (ft^3/d)	.2832	cubic meter per day
cubic foot per day per mile [$(ft^3/d)/mi$]	0.177	cubic meter per day per kilometer

Temperature in degrees Celsius (°C) can be converted to degrees Fahrenheit (°F) as follows:

$$^{o}F = 1.8 \times {}^{o}C + 32$$

Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

HYDROGEOLOGY AND GEOCHEMISTRY IN BEAR CREEK AND UNION VALLEYS, NEAR OAK RIDGE, TENNESSEE

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ABSTRACT

Ground-water flow in Bear Creek Valley, which contains the Y-12 Plant complex (a nuclear-component production facility) and numerous hazardous-waste disposal areas, is primarily from the ridges toward the main streams on the valley floor. The main streams in the valley are incised into the Maynardville Limestone. The ground-water flow system, recharged primarily on Pine and Chestnut Ridges, discharges to the Maynardville Limestone and ultimately to streams flowing on the Maynardville. Contaminants reaching the Maynardville could be transported by ground water along the strike of the geologic formation or out of the valley by streams.

Ground-water flow in the valley is primarily normal to strike; however, short flow paths along strike (down the valley) are controlled by closely spaced ephemeral streams that are normal to strike. Ground-water flow along strike is also facilitated by localized zones of more intense fracturing or solution cavities. The flow system is generally continuous across the geologic formations and at depth. Results of geochemical models for water in the Rome Formation, Maynardville Limestone, and Copper Ridge Dolomite indicate that more of the ground water flowing to the Maynardville Limestone is from Chestnut Ridge than from Pine Ridge. Four flow zones in the valley were distinguished by using potentiometric data: 0 to 50 feet below land surface, 50 to 100 feet, 100 to 400 feet, and deeper than 400 feet.

Two zones of water chemistry, from 0 to 50 feet and 50 to 500 feet below land surface, were distinguished using geochemical data. Although the chemistry of both zones is dominated by calcium and bicarbonate ions, the deeper zone is distinguished by chemical evolution to sodium and bicarbonate dominance along ground-water flow paths. Areas of elevated concentrations of dissolved solids (as much as 15,000 milligrams per liter in the deep zone and 20,000 milligrams per liter in the shallow zone) indicate contamination from waste-disposal sites.

Hydraulic conductivity used in the digital model for the geologic formations ranged from 0.3 to 0.0016 foot per day for the upper 400 feet of strata. A value of 0.000078 foot per day was used for the part of all formations deeper than 400 feet below land surface.

Areal recharge provided all the incoming water to the modeled system, although problems during calibration in matching some head gradients and results of sensitivity analyses may indicate a need for a source of ground-water underflow from Pine Ridge. All of the discharge from the system is to the main streams, and 23 percent of that discharge is to the normal-to-strike tributaries. Therefore, the streams are probably the primary recipients of any contaminants in the ground water. The most likely area of potential transport of contaminants beyond the Oak Ridge Reservation property through the ground-water system is from the eastern end of the Y-12 Plant complex, where

ground water from the East Fork Poplar Creek basin may flow into the Scarboro Creek basin.

INTRODUCTION

The Y-12 Plant, a nuclear-component production facility, occupies about 450 acres near the boundary of the U.S. Department of Energy's (DOE) Oak Ridge Reservation (ORR) on the east end of Bear Creek Valley. Numerous hazardous-waste disposal and storage sites are located within the Y-12 Plant complex, and four major disposal sites are situated in the valley: the S-3 ponds, the Oil Landfarm, the Bear Creek Burial Grounds, and New Hope Pond. The U.S. Geological Survey (USGS), in cooperation with the DOE, conducted a study of the hydrogeology of Bear Creek Valley, which lies within the Reservation. The area of investigation was extended to include Union Valley, outside the ORR boundary, because it is a geographic extension of Bear Creek Valley (fig. 1). The study area is about 15 mi^2.

PURPOSE AND SCOPE

The purpose of this investigation was to formulate an understanding of the ground-water flow system and geochemistry in the vicinity of the Y-12 Plant, and to determine potential pathways of contaminant migration resulting from Plant effluent and land disposal of wastes and hazardous material. The objectives were to (1) quantify the flow, quality, and interaction of surface water with the ground-water system; (2) determine hydraulic characteristics of geologic units; (3) develop a concept of the valley-wide ground-water flow system; (4) quantify the components of the water budget; and (5) simulate the dynamics of the flow system and identify potential directions (or general pathways) of contaminant migration. This report summarizes the results of the investigation and describes the hydrogeology of Bear Creek Valley.

A minimal amount of new data was collected specifically for this investigation. Rather, existing data from previous investigations and information collected in concurrent local investigations by Martin Marietta Energy Systems, Inc. (MMES) and their contractors were used as much as possible.

APPROACH

The investigation was conducted in several phases to provide the geologic and hydrologic information necessary to describe the ground-water flow system. Data and results from most of the phases are published in separate reports.

Stream discharge and specific conductance of water were measured along all streams and at springs in Bear Creek and Union Valleys, and on Chestnut Ridge during the period February 15 through April 9, 1984 (Evaldi, 1984). Discharge and specific conductance were measured along Bear Creek again on August 13, 1985 (Evaldi, 1986). The measurements in April 1984 were done during high base flow, and those in August 1985 were done at low base flow. Thirty-four of the sites measured during the April 1984 reconnaissance were selected for more intensive water-quality analyses, and sampling was done on April 13 and 14, 1984, during high base flow and again on September 26 and 27, 1984, during low base flow (Pulliam, 1985a, b).

During the course of the investigation, available information on 547 wells and test borings was compiled in a computer data base to produce geologic and hydrologic maps, and to formulate concepts of the flow system. About 400 test borings and wells had been drilled by the beginning of the investigation. The remaining borings and wells were drilled subsequently by the USGS (Bailey and Withington, 1988) and by contractors to MMES.

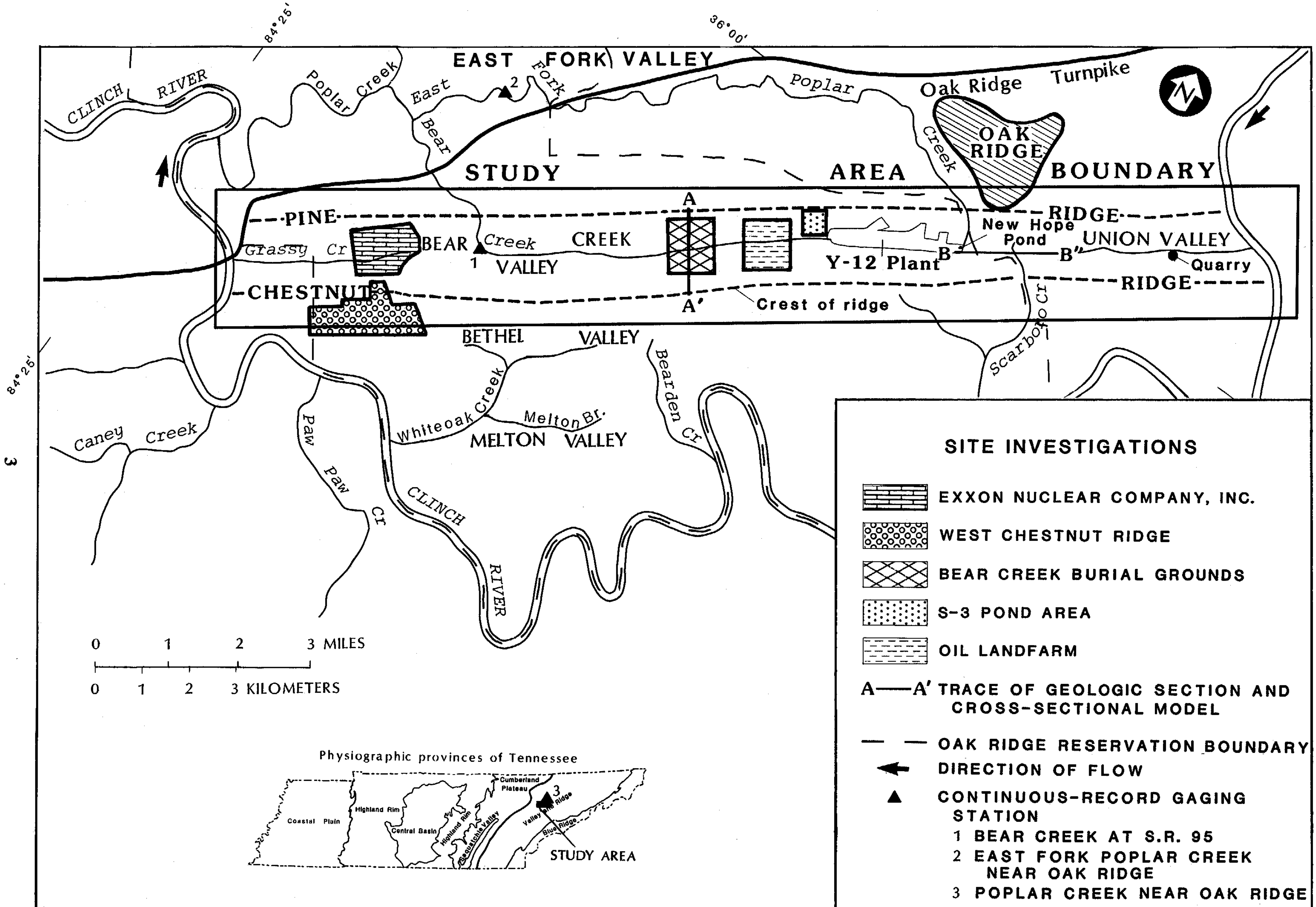


Figure 1.--Location of study area.

Geologic maps delineating formations in the Conasauga Group and thickness of the regolith in the valley and on the ridges for all of Bear Creek and Union Valleys were prepared (Hoos and Bailey, 1986), using available geologic data and maps from local studies. The purpose of the maps was to provide more detailed information for constructing ground-water flow models.

A preliminary cross-sectional ground-water flow model (A-A′, fig. 1) was constructed to determine where additional data were needed to simulate flow, and to study cross-valley flow at depth and areal recharge distribution (Bailey, 1988). Hydraulic information was particularly lacking for depths greater than 100 feet below land surface and along the hydrologic divides of the ridges. This lack of information was persistent throughout the valley. Additional deep wells along the line of cross section were subsequently completed on the ridges and at depths of as much as 600 feet below land surface in the valley. Information from these wells was used to revise the cross-sectional model. Results of the cross-sectional model are used in this report to demonstrate flow patterns at depth.

Additional wells were installed by the USGS at nine sites around the perimeter of the valley. These wells provided information necessary to define boundary conditions for the flow models and supplemented geologic and hydrologic data from more localized studies within the valley. Three wells were drilled at most of the sites: one shallow water-table well, one to a depth of about 100 feet, and one to a depth of about 400 feet. Geophysical logging was completed in six of the deepest wells (Bailey and Withington, 1988).

Hydraulic-conductivity values from 338 single-well aquifer tests were analyzed statistically to determine representative hydraulic conductivities for each geologic unit. A cross-sectional ground-water flow and regression model was used to further refine the conductivity values for use in the three-dimensional flow model (Connell and Bailey, 1989).

This report describes the ground-water flow system. Information from the preliminary phases of the investigation was used to conceptualize the system and to construct a three-dimensional ground-water flow model. The model was used to simulate the system, test the estimates of hydraulic properties, determine the importance to the system of ground water and stream interaction, quantify the components of the water budget, and demonstrate potential directions of contaminant migration. The geochemical nature of the ground water was investigated to verify concepts of the flow system derived from potentiometric data. The chemical evolution of ground water is represented in this report by Piper diagrams, maps of the distribution of chemical constituents, and geochemical models. Chemical analyses from 142 wells were used in the geochemical interpretations; 19 were used for the geochemical modeling. Nearly all of the ground-water chemical data used were collected and analyzed by contractors to MMES. However, 19 of the wells drilled for this investigation were sampled by USGS staff for major chemical constituents. Temperature, specific conductance, pH, bicarbonate, and carbonate, were measured at the wellhead during sample collection.

PREVIOUS INVESTIGATIONS

Several site-specific studies have been done in Bear Creek Valley and on Pine and Chestnut Ridges (fig. 1). Well and borehole locations, ground-water levels, hydraulic-conductivity data, quality of ground and surface water, and local geologic and hydrologic interpretations from these studies were used to formulate preliminary interpretations of the flow system and were supplemented by data collected during this study.

An extensive hydrogeologic investigation of Pine Ridge and Bear Creek Valley in the Grassy Creek watershed (Exxon Nuclear Company, Inc., 1978) provided the only geologic and hydrologic data available for that segment of the valley. Ketelle and Huff (1984) and Woodward-Clyde Consultants (1984) investigated the hydrogeology of a segment of Chestnut Ridge that is partly in the same watershed as the Exxon study. Bechtel National, Inc. (1984a-f) produced a series of data and interpretive reports on the Oil Landfarm and Bear Creek Burial Ground areas. Data from drilling, water-level measurements, and sampling for water quality and detection of contaminants were available from more recent studies conducted in the waste-disposal areas by Geraghty & Miller, Inc. (1985a, b, 1987). A study within the Y-12 Plant (Rothschild and others, 1984) provided the only geologic and hydrologic information in that segment of the valley. Hydrologic and hydrochemical assessments of shallow ground water were also done in the vicinity of the Y-12 Plant (Haase and others, 1987a, b). Unpublished ground-water level and quality data collected by contractors and by the staff of MMES were also made available for use in this investigation.

Investigators have found five primary types of contaminants in ground water of the Y-12 area: nitrates, heavy metals, radioactivity, volatile organic compounds (VOC), and high dissolved solids (Geraghty and Miller, Inc., 1985a; Haase and others, 1987a). Three principal disposal areas, the S-3 ponds, the Oil Landfarm, and the Burial Grounds, contain most of the contaminants near the Y-12 Plant. Geraghty & Miller, Inc. (1985a) describe the history and contents of these disposal areas. New Hope Pond, at the eastern end of the Y-12 Plant, is a settling basin for Y-12 Plant effluent that flows to East Fork Poplar Creek. The pond is within 1,000 feet of the ORR perimeter.

DESCRIPTION OF STUDY AREA

GEOGRAPHY AND CLIMATE

The ORR is near the northwestern edge of the Valley and Ridge physiographic province (fig. 1), which is characterized by repeating sequences of elongate ridges and intervening valleys, all trending northeast-southwest (Miller, 1974, p. 3). Bear Creek and Union Valleys are narrow, less than one-half mile wide, and their topography is generally flat to rolling; land-surface elevations range from 750 to 1,000 feet above sea level. Pine Ridge rises steeply to about 300 feet above the valley floor and is heavily wooded. Chestnut Ridge is not as high or as steep as Pine Ridge, but it is also wooded on the slope adjacent to Bear Creek Valley.

Mean annual temperature in the area is 57 °F (14 °C), and mean annual precipitation is 54 inches, calculated for the period 1956 to 1985 (National Oceanic and Atmospheric Administration, 1985, p. 4B).

DRAINAGE FEATURES

Stream courses in the Valley and Ridge are controlled by geologic structure and lithology. Major streams flow parallel to the axes of the valleys, which are underlain by more easily eroded rock units. The drainage in the area forms a trellis pattern because tributaries to the major drainage are influenced by rock fractures perpendicular to the trend of the valleys (Miller, 1974, p. 3).

The Clinch River is the major drain for the area and surrounds three sides of the ORR (fig. 1). Flow rates and water levels of the river are controlled on the west end of Bear Creek Valley by Watts Bar Dam and on the east end of Union Valley by Melton Hill Dam. Average discharge for 37 years of record in the vicinity of Melton Hill Dam is 4,592 ft^3/s (Lowery and others, 1987, p. 114). Average elevation of the river surface during October is 794 feet above sea level at the west end of the study area, and 740 feet above sea level on the east end (William Feltz, Tennessee Valley Authority, written commun., 1986).

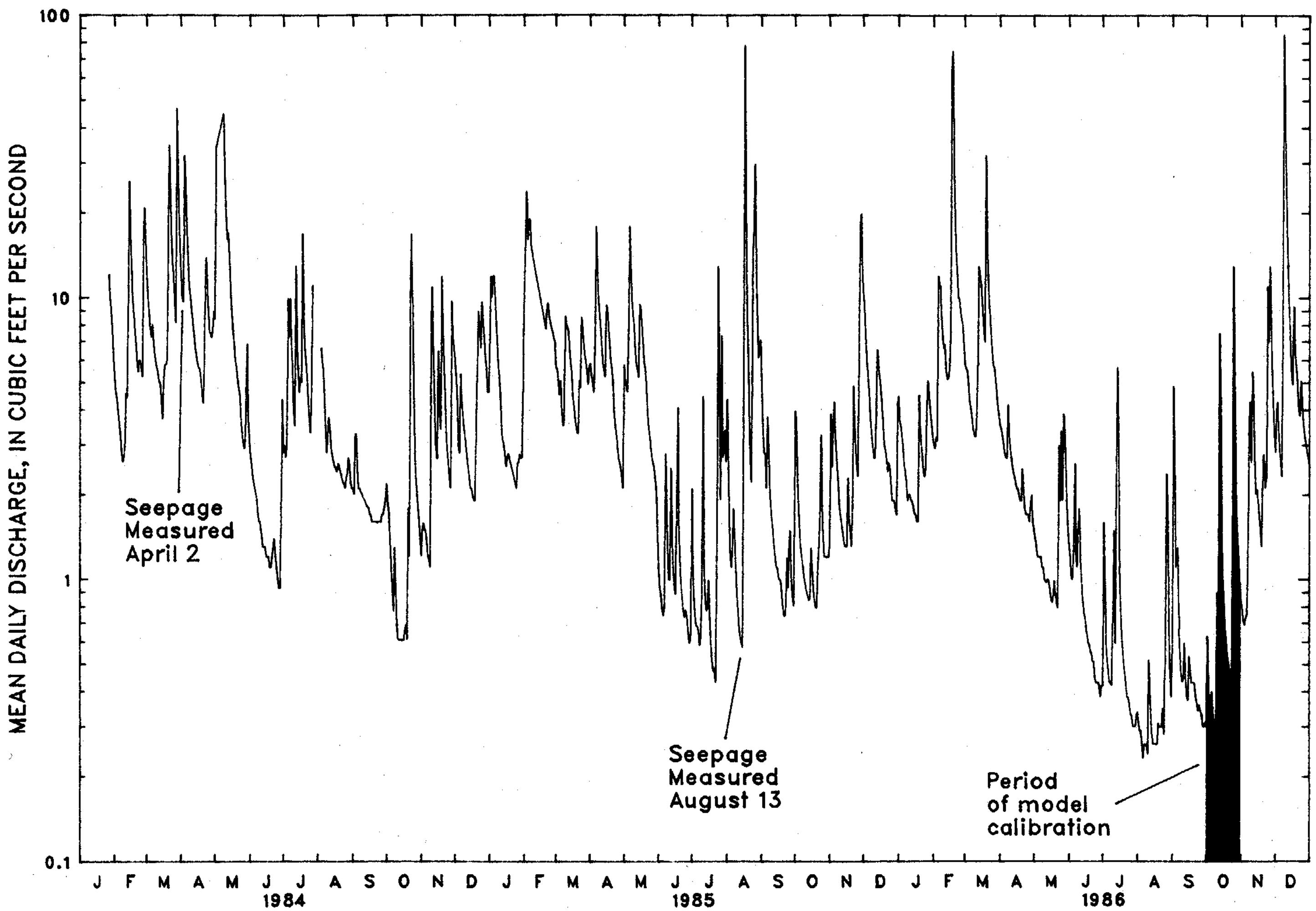
The divides between the watersheds of East Fork Poplar Creek, Bear Creek, Grassy Creek, and Scarboro Creek are formed by slightly higher areas in the rolling terrain of the valley floor. The Y-12 Plant complex is in the East Fork Poplar Creek watershed.

The natural headwaters of East Fork Poplar Creek are ditched or buried beneath the Y-12 Plant complex in Bear Creek Valley. The natural stream flowed northeasterly along the axis of the valley. Flow is released to the channel downstream of the Y-12 Plant through New Hope Pond. The stream channel turns 90 degrees to flow through a gap in Pine Ridge, and through the city of Oak Ridge. Beyond Oak Ridge, the channel in East Fork Valley parallels the headwater drainage, but flow is in the opposite direction of flow in Bear Creek Valley. East Fork Poplar Creek drains into Poplar Creek just upstream of the confluence of Poplar Creek and the Clinch River. Flow in East Fork Poplar Creek is maintained year round by effluent from the Y-12 Plant, which may be as much as 20 ft^3/s. Under natural conditions the headwaters of the stream would be dry much of the time. A municipal sewage treatment plant just downstream from the city adds as much as 10 ft^3/s to the streamflow. The unadjusted average discharge of East Fork Poplar Creek at the continuous-record gage (fig. 1) over 26 years of record (1960 to 1986) is 50.7 ft^3/s (Lowery and others, 1987, p. 125).

The headwaters of Bear Creek are near the S-3 Ponds. Bear Creek flows southwesterly along the axis of the valley through Pine Ridge, and drains into East Fork Poplar Creek. Small tributaries to Bear Creek are in a regularly spaced rectangular pattern draining Pine Ridge. Few surface tributaries drain Chestnut Ridge; the drainage is primarily subsurface and runoff reaches Bear Creek through numerous springs along the base of the ridge. Intermittent measurements of streamflow in Bear Creek were made during the period April 1959 to June 1964, and a continuous-record gage (fig. 1) was operated during this study at a weir that is the control from 0 to 1.46 feet of stage. Using the most recent rating for the station, a stage of 1.46 feet corresponds to a stream discharge of 48 ft^3/s. Ratings have been developed for discharge greater than 48 ft^3/s using streamflow measurements during times of flow above the weir. The range of mean-daily discharge, from January 1984 through December 1986, is 0.23 to 86 ft^3/s (fig. 2). Maximum instantaneous discharge during this period was 145 ft^3/s on February 17, 1986, and the minimum recorded discharge was 0.22 ft^3/s on August 5, 6, and 10, 1986 (Lowery and others, 1987, p. 126).

Grassy Creek drains the western end of Bear Creek Valley and flows into the Clinch River. There are no continuous records of stream discharge for Grassy Creek, but an average discharge of 3 ft^3/s at the confluence with the Clinch River has been estimated (Exxon Nuclear Company, Inc., 1978, p. 3.4-3).

The headwaters of Scarboro Creek are on Pine Ridge and the creek cuts through all the geologic units of Bear Creek Valley and Chestnut Ridge as it flows southeast to the Clinch River. The divide between East Fork Poplar Creek and Scarboro Creek was investigated during the study to determine if the divide also extends to the deep ground-water flow system. Flow in the unnamed creek in Union Valley was measured in April 1984 (Evaldi, 1984). The stream had little to no flow upstream from the quarry (fig. 1) and flow downstream was sustained by water pumped from the quarry. Flow at the mouth of the stream was 0.32 ft^3/s.



Data for January, 1984, through February, 1985, from Y–12 National Pollution Discharge Elimination System; data for March, 1985, through December, 1986, from U.S. Geological Survey stream gaging station

Figure 2.––Mean daily discharge of Bear Creek at Highway 95, 1984–86.

GEOLOGY

Bear Creek and Union Valleys and adjacent ridges are underlain by rocks of Cambrian and Ordovician age that strike north 56 degrees east. The dip of the rocks is from 30 to 70 degrees southeast; average dip is about 45 degrees. Bedrock is overlain by clay-rich regolith, which often retains relict structure of the bedrock and contains rock fragments.

BEDROCK

Pine Ridge is underlain by interbedded sandstone, siltstone, and shale of the Rome Formation; Bear Creek Valley is underlain by calcareous shale and limestone of the Conasauga Group; and Chestnut Ridge is underlain by massive, siliceous dolomite of the Knox Group and contains solution and karst features (McMaster, 1963, p. 6, 8, 10). The same geologic sequence is repeated in Melton Valley (fig. 1) and Poplar Creek Valley by the thrust faulting that formed the Valley and Ridge terrain, and for this reason, some data and hydrologic interpretations from these two valleys were applied in this investigation to interpretation of the hydrogeology of Bear Creek Valley. A thrust fault in the Rome Formation and subparallel to Pine Ridge (fig. 3) is part of the Whiteoak Mountain fault system (McMaster, 1963, p. 19).

Formations in the Conasauga Group (fig. 3), from oldest to youngest, are the Pumpkin Valley Shale, Rutledge Limestone and Rogersville Shale (regarded as one unit for this study), Maryville Limestone, Nolichucky Shale, and Maynardville Limestone (Rodgers, 1953, p. 47). The Maynardville Limestone is the formation of greatest hydrologic interest in the study because of ground-water flow through fractures and cavities.

Formations in the Knox Group (fig. 3), from oldest to youngest are the Copper Ridge Dolomite, the Chepultepec Dolomite, and the undifferentiated upper part of the Knox Group. The Chickamauga Limestone overlying the Knox Group is undivided for this investigation.

REGOLITH

The regolith, which consists of soil and weathered rock, ranges from 0 to 80 feet in thickness and overlies the bedrock except where rock crops out in stream channels. Regolith tends to be thickest on the ridges and thins into the valley; several reaches of Bear Creek flow on bedrock.

HYDROGEOLOGY

HYDRAULIC CHARACTERISTICS OF REGOLITH AND BEDROCK

Hydraulic-conductivity values range over several orders of magnitude (from 0.00002 to 136 ft/d, Connell and Bailey, 1989, p. 9) due to the low permeability formations, which have secondary permeability along bedding planes, fractures, and solution cavities. Available hydraulic-conductivity values from 338 single-well aquifer tests were analyzed statistically to determine the median and variation of hydraulic conductivity within each geologic unit. These analyses showed that (1) conductivity differs substantially among geologic units, and (2) the conductivity of regolith is greater than that of bedrock; however, the difference between regolith and bedrock is not significant except for the Nolichucky Shale (Connell and Bailey, 1989, p. 12-14). The median values for each formation were further refined using a ground-water flow and regression model (Connell and Bailey, 1989). Model results were little affected by treating the regolith and bedrock separately in the Nolichucky Shale; therefore, only hydraulic-conductivity values determined in this phase of the investigation for bedrock were used as initial estimates of hydraulic conductivity (table 1) for each formation and each layer (discussed later) in the three-dimensional model.

Table 1.—*Initial hydraulic-conductivity values for the digital flow model* (Connell and Bailey, 1989, p. 25)

Geologic unit (oldest to youngest)	Hydraulic conductivity, in feet per day		
	Layers 1 and 2	Layer 3	Layer 4
Rome Formation	0.30	0.03	0.000078
Pumpkin Valley Shale	.016	.0016	.000078
Rutledge Limestone and Rogersville Shale	.037	.0037	.000078
Maryville Limestone	.034	.0034	.000078
Nolichucky Shale	.059	.0059	.000078
Maynardville Limestone	.039	.0039	.000078
Copper Ridge Dolomite	.031	.0031	.000078

GROUND- AND SURFACE-WATER INTERACTION

The hydraulic connection between the ground water and surface water can be determined using flow-duration curves and base-flow measurements. Flow-duration curves were used to compare basin characteristics of Bear Creek, East Fork Poplar Creek, and Poplar Creek. Seepage characteristics of geologic units were determined in Bear and Scarboro Creeks using base-flow measurements.

Hydrologic and geologic characteristics of a drainage basin and comparative characteristics of basins can be determined by the shape of a flow-duration curve. A steep slope indicates a highly variable stream whose flow is mainly from direct runoff. A steep slope at the lower end of the curve indicates a negligible amount of perennial storage in the basin; a flat slope indicates a large amount of storage (Searcy, 1959, p. 22).

McMaster (1967) constructed flow-duration curves from discharge data collected at gaging stations on Bear Creek, East Fork Poplar Creek, and Poplar Creek. The curves for mean daily flow were converted to mean daily flow per square mile of drainage basin for each of the creeks (fig. 4) so that unit values could be compared.

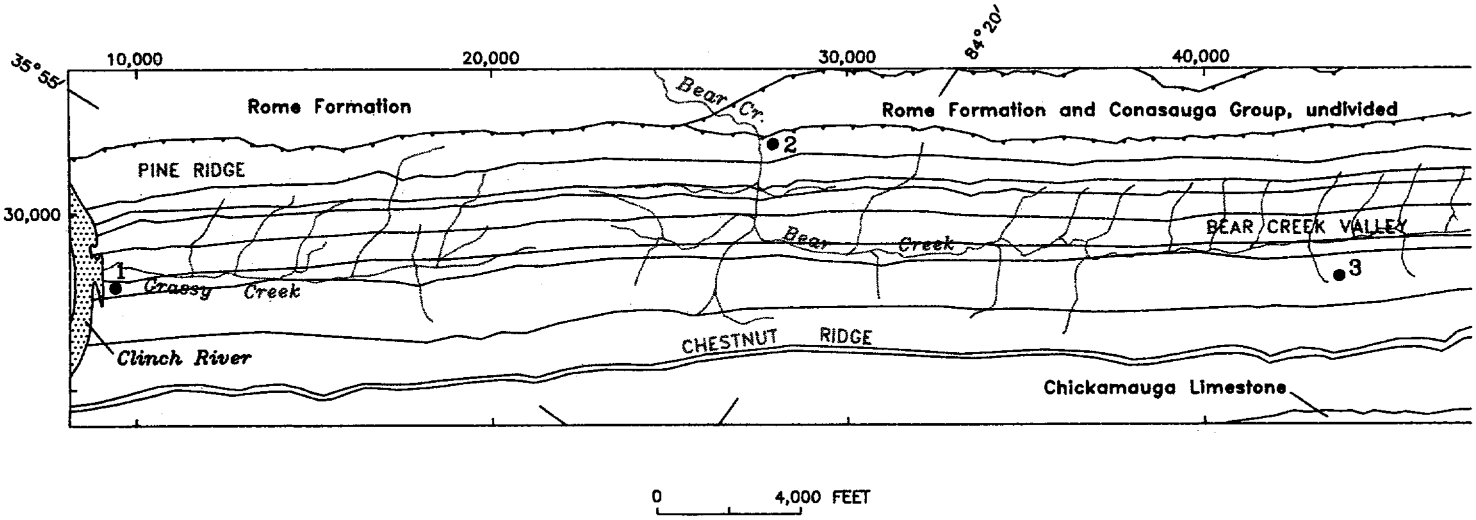
The slopes of the curves at high flows (above 60 percent flow duration) are nearly the same, indicating that the climate, physiography, and plant cover are similar (or have similar combined effects on runoff) in all the basins (Searcy, 1959, p. 24).

The distribution of low flow, indicating the effects of geology on ground-water runoff (Searcy, 1959, p. 24), is nearly identical for Bear Creek and Poplar Creek, and little storage capacity in the basins is indicated. Both streams flow primarily on the Conasauga Group and drain areas of the Rome Formation and Knox Group.

The lower part of the duration curve for East Fork Poplar Creek indicates more storage capacity in the basin than in Bear Creek or Poplar Creek basins. East Fork Poplar Creek flows on rocks of the Chickamauga Limestone, although some drainage is from the Knox Group and the headwaters drain the Conasauga Group and Rome Formation.

Base-flow measurements were made for Bear Creek (fig. 2) and Scarboro Creek at periods of high base flow (Evaldi, 1984) and for Bear Creek at low base flow (Evaldi, 1986). Analysis of data from these investigations indicated that both are gaining streams even though they lose water to the ground-water system along some reaches.

Relative differences in seepage characteristics of each geologic unit could be determined from base-flow measurements on Scarboro Creek, because the creek cuts through



EXPLANATION

—— GEOLOGIC CONTACT

—— THRUST FAULT

● WELL LOCATION FOR WATER CHEMISTRY:

1 GW-212, GW-214, GW-238
2 GW-209, GW-210, GW-211
3 GW-165, GW-166
4 GW-206. GW-209, GW-208
5 GW-167, GW-168, GW-239
6 GW-170
7 GW-171, GW-172, GW-230

CHICKAMAUGA LIMESTONE — ORDOVICIAN

KNOX GROUP:
- UPPER KNOX GROUP, UNDIVIDED — ORDOVICIAN
- CHEPULTEPEC DOLOMITE — ORDOVICIAN
- COPPER RIDGE DOLOMITE — CAMBRIAN

CONASAUGA GROUP — CAMBRIAN:
- MAYNARDVILLE LIMESTONE
- NOLICHUCKY SHALE
- MARYVILLE LIMESTONE
- ROGERSVILLE SHALE AND RUTLEDGE LIMESTONE

PUMPKIN VALLEY SHALE

ROME FORMATION

Note: Grid coordinate system unique to this map (S-16A)

Figure 3.--Bedrock geology.

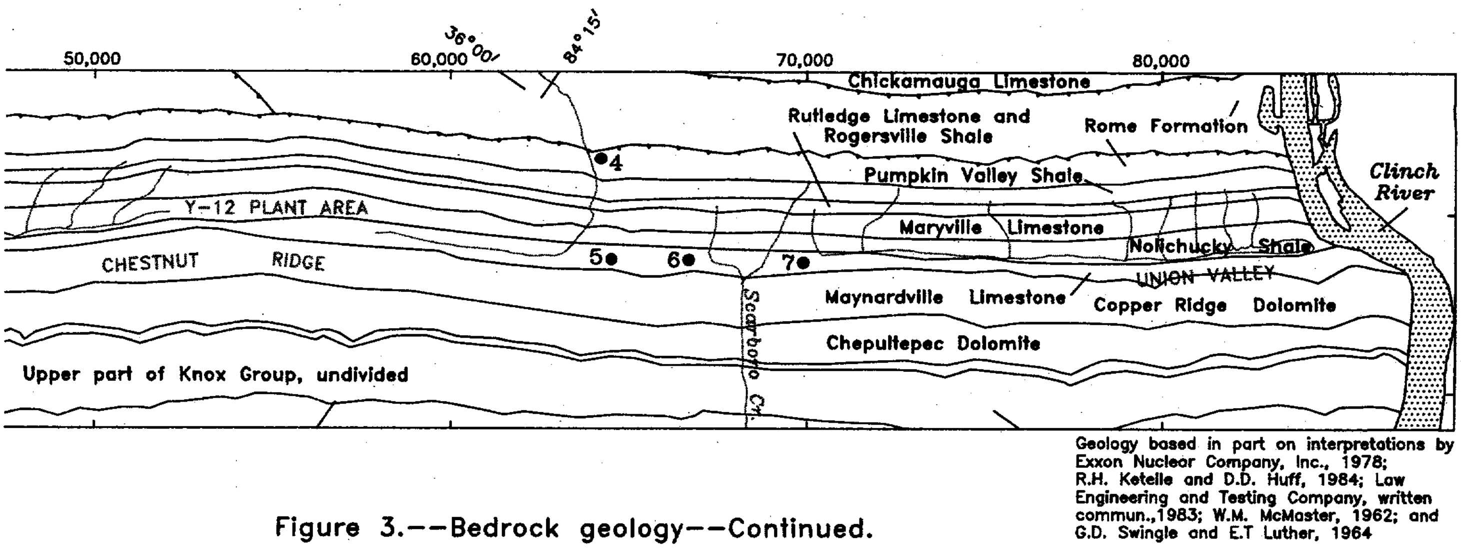


Figure 3.—Bedrock geology—Continued.

the entire geologic section (fig. 3). Differences calculated between consecutive discharge measurements along the stream show little gain or loss of ground water in the reaches of the stream in the Rome Formation and the Conasauga Group, but significant gains and losses associated with the rocks containing fractures and solution features in the Knox Group and in the Chickamauga Limestone (fig. 5). Over the entire reach, however, Scarboro Creek has a net gain of ground water.

Although Bear Creek loses flow along some reaches (fig. 6), the flow is regained because the water leaves and enters the bedrock channel through fractures and solution cavities in the underlying bedrock. Bear Creek flows mainly subparallel to the strike of the Maynardville Limestone, which has numerous solution cavities. The total ground-water gain for Bear Creek measured April 2, 1984, was 8.4 ft^3/s, which is an overall seepage rate of about 1 ft^3/s per mile of stream channel. Measurements made during low base flow, August 13, 1985, indicated an overall loss of 0.004 ft^3/s of streamflow to the ground-water system. However, the difference between total streamflow and total inflow from ground water is so slight that this difference could be attributed to measurement error rather than to any actual overall loss.

RECHARGE

Recharge to the water table in regolith or exposed bedrock is from precipitation. Ground water in the deeper bedrock is recharged by water percolating through the regolith. Several methods were used to calculate recharge rate,

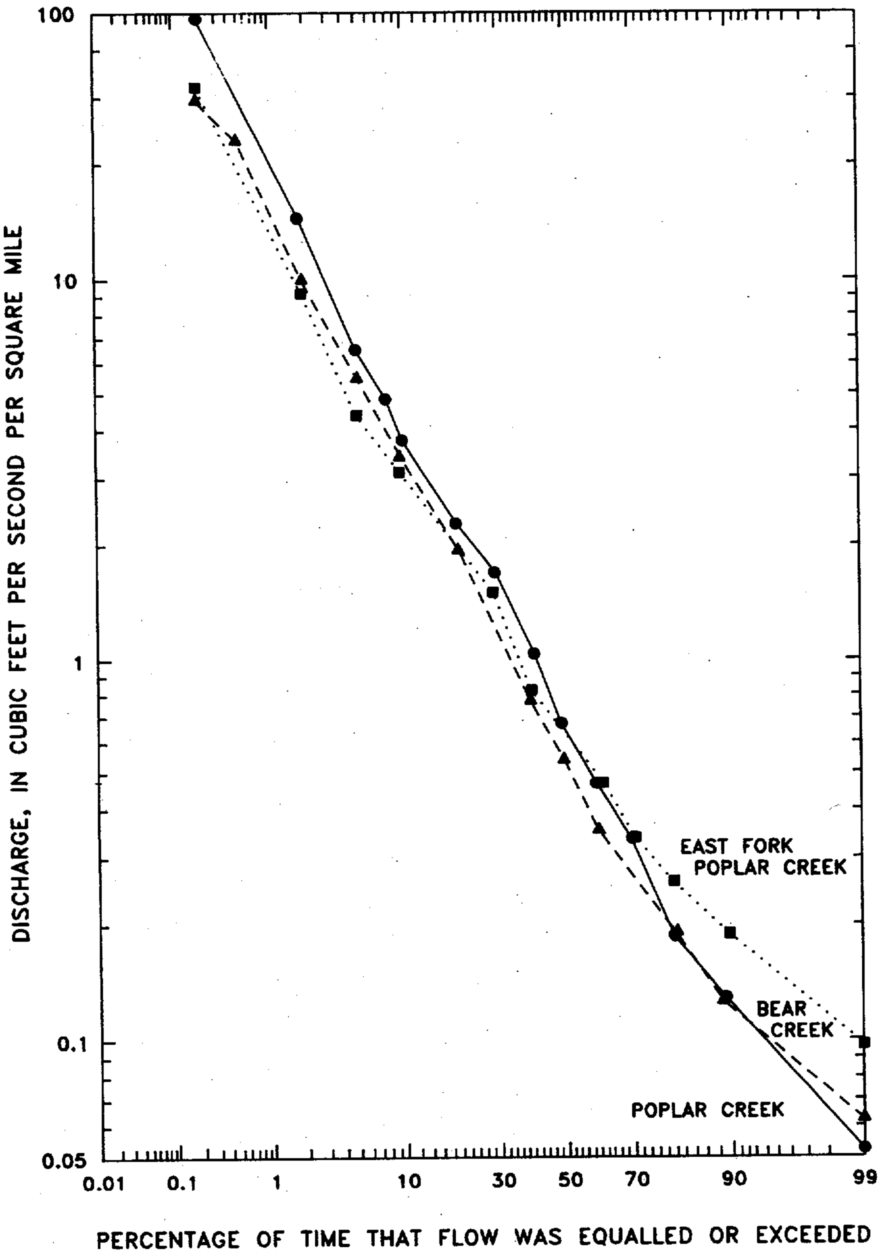


Figure 4.--Flow duration for Bear Creek, East Fork Poplar Creek, and Poplar Creek. (Adjusted to 1936-60 period and for wastewater discharges; modified from W.M. McMasters, 1967.)

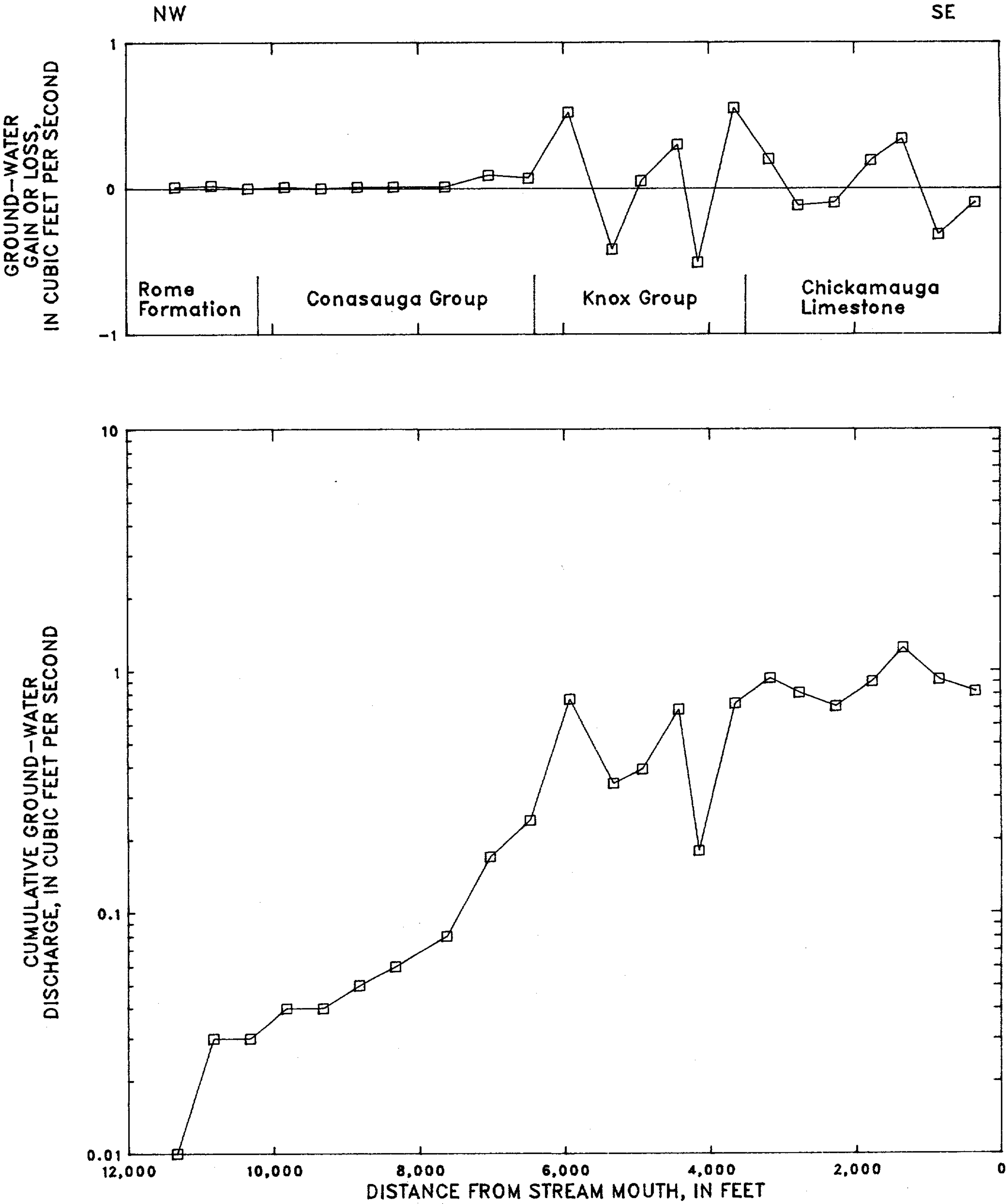


Figure 5.--Ground-water gains and losses and cumulative ground-water discharge along Scarboro Creek, March 10, 1984.

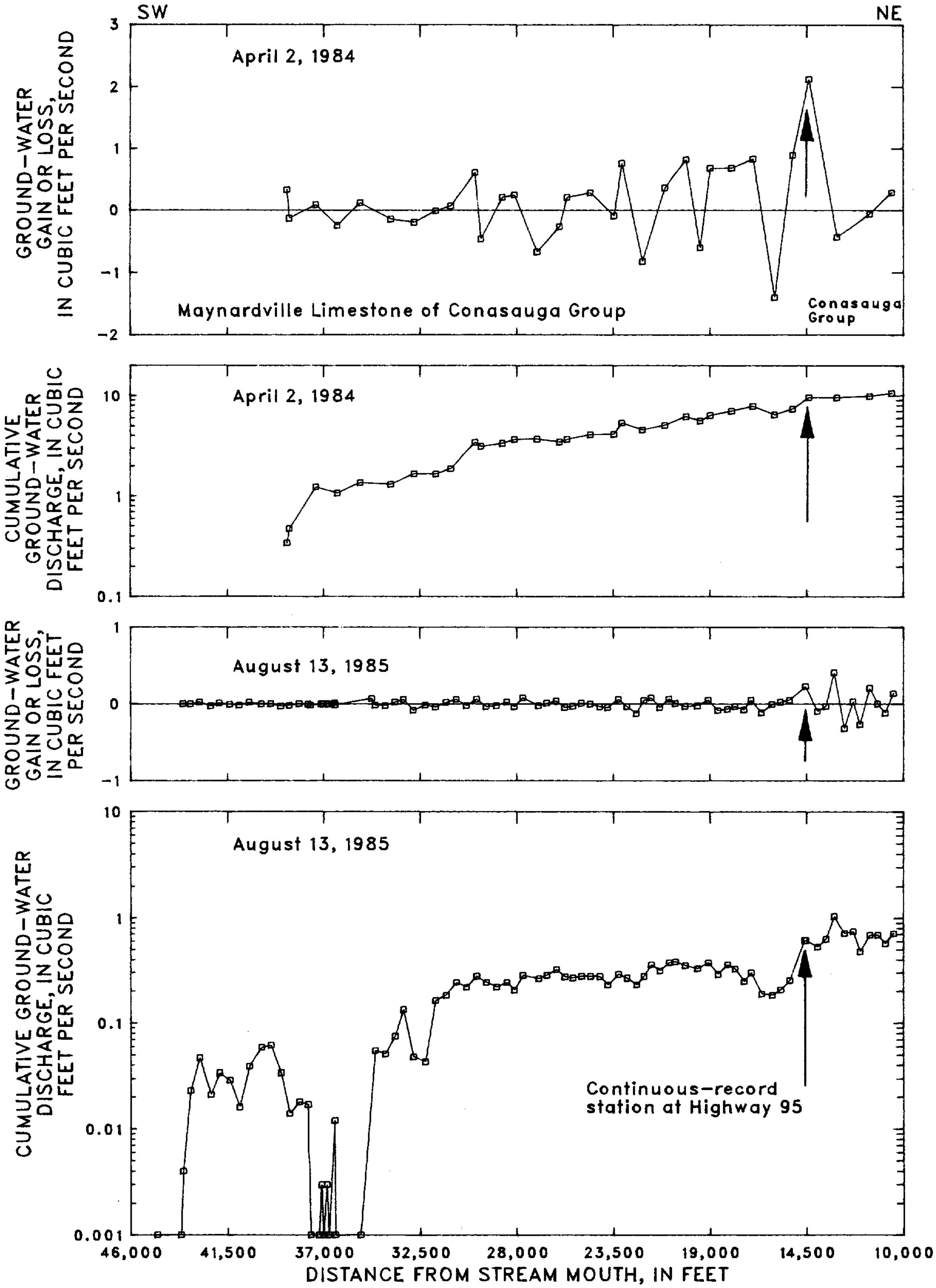


Figure 6.--Ground-water gains and losses and cumulative ground-water discharge along Bear Creek.

and a wide range of rates resulted from the various methods.

The period of continuous record for flow in Bear Creek is insufficient to analyze for recharge; however, Poplar Creek, which drains a similar geologic area, has continuous streamflow records for 23 years (1961-83). It was assumed that the percentage of precipitation that is recharge to the ground-water system in the Poplar Creek basin could be applicable to the Bear Creek basin because of the similar geology, and because analysis of flow-duration curves for Bear Creek and Poplar Creek (see section "*Ground- and Surface-Water Interaction*") indicates that the basins have similar recharge characteristics. A hydrograph-separation technique (Rorabaugh, 1964; Daniel, 1976) was used to estimate annual recharge in the Poplar Creek basin (R.D. Evaldi, U.S. Geological Survey, written commun., 1984). Mean annual recharge for 23 years of record was estimated to be 14 in/yr, which is 25 percent of mean annual precipitation.

The same hydrograph-separation technique was applied to records from selected years that were determined to be typical (1984), wet (1973), and dry (1985). Rainfall during the typical year was the median value for all the years of record, and rainfall during the wet and dry years were the values from years that represented the extremes during the years of record (A.B. Hoos, U.S. Geological Survey, written commun., 1986). The estimates of recharge were 15, 20, and 12 percent of annual precipitation, respectively, for typical, wet, and dry years. Recharge calculated from those percentages ranged from 6 to 15 in/yr using the actual precipitation for each year calculated (1973, 1984, 1985), or from 7 to 11 in/yr using mean annual precipitation.

The areal distribution of recharge is unknown, but soils developed on the different geologic formations have different capacities for retaining precipitation and percolating water downward to the water table. Soil on Pine Ridge and on the valley floor has a slow infiltration rate, and soil on Chestnut Ridge has a moderate infiltration rate (U.S. Department of Agriculture, 1981, p. 58, 162-163). The regolith, which is thicker on the ridges, retains water that slowly recharges underlying formations, and the sinkholes on Chestnut Ridge also encourage infiltration rather than runoff.

The distribution of recharge was one of the variables investigated in a cross-sectional finite-difference model across the valley at the Burial Grounds (Bailey, 1988), and in the regression model (Connell and Bailey, 1989). Most of the recharge calculated by the cross-sectional model is on the ridges; 25 in/yr on Pine Ridge, and 20 in/yr on Chestnut Ridge. The net recharge estimated for the whole modeled area was 10 in/yr (Bailey, 1988). Further revisions to the cross-sectional model, based on water levels from additional wells, produced an overall average recharge of about 14 in/yr, but the pattern of recharge and discharge distribution remained the same. The formations between Pine Ridge and Bear Creek are primarily discharging, although both recharge and discharge occur. Most of the discharge from the system is through the Maynardville Limestone to Bear Creek.

Estimates of recharge from the hydrograph-separation technique and cross-sectional modeling are high compared to estimates from an areal flow model in nearby Melton Valley (Tucci, 1986) and from recharge calculations by Moore (1988). Tucci (1986, p. 11) calibrated a preliminary, areal ground-water flow model using a ground-water recharge rate of 3.2 in/yr. Calculations by Moore (1988, p. 33, 85) using data from Chestnut Ridge, Melton and Bethel Valleys resulted in subsurface recharge estimates of 2.6 to 5.2 in/yr. Moore calculated that 90 to 95 percent of the recharge that enters the subsurface, flows through the "stormflow-zone," which is above the water table, and never enters the ground-water system. The remaining 5 to 10 percent of the subsurface recharge enters the ground-water system. Moore extended his theory of a stormflow zone to Bear Creek Valley and calculated a ground-water recharge rate of 2.95 in/yr (G.K. Moore, University of Tennessee, Knoxville, written commun., 1989).

WATER-LEVEL FLUCTUATIONS

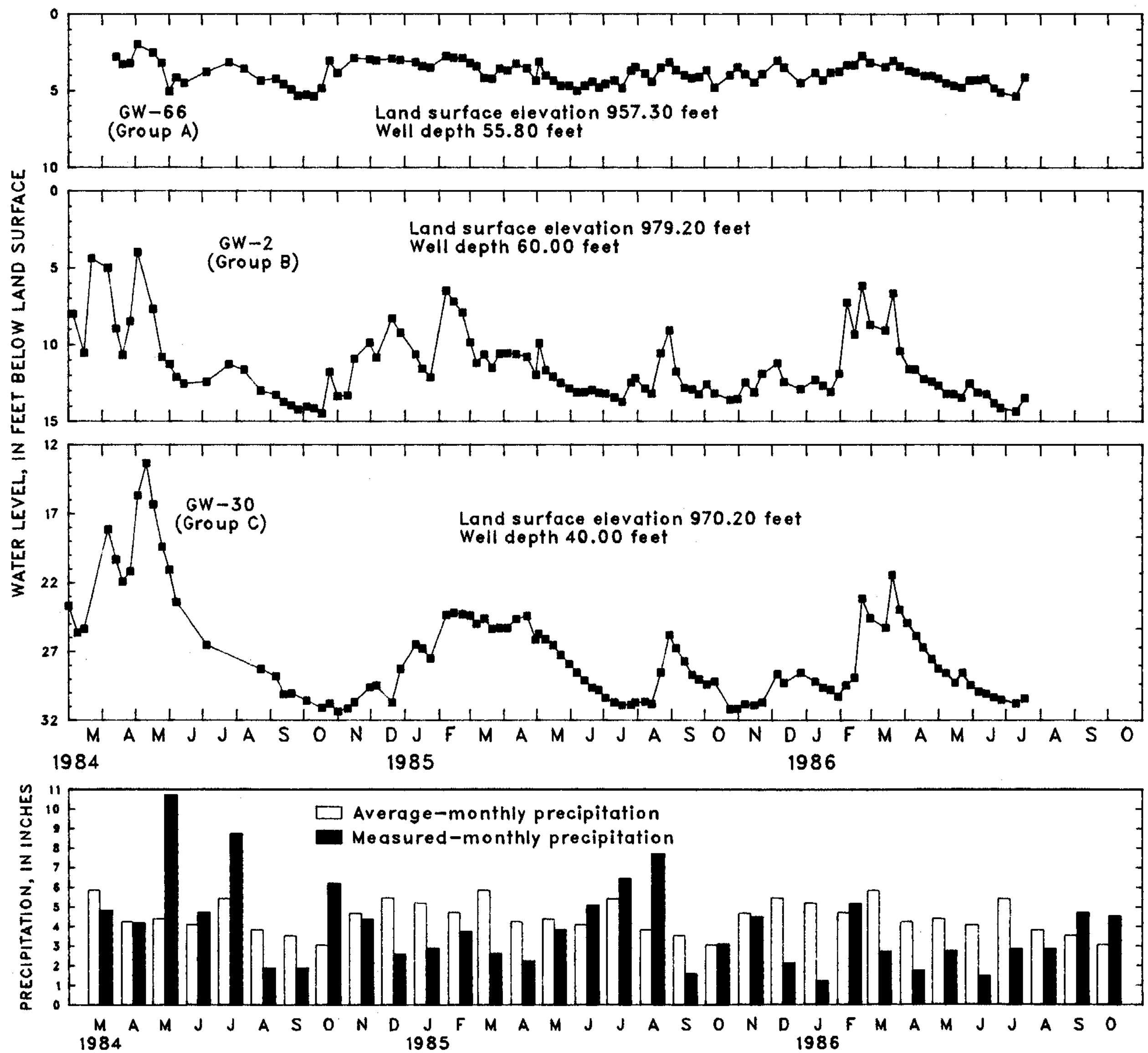
Natural seasonal fluctuations of the water table are related to seasonal changes in precipitation, evapotranspiration, and thus, to changes in ground-water recharge. Ground-water levels, which are normally highest during the spring months due to high precipitation and low evapotranspiration during the winter months, recede during the summer in response to low precipitation and high evapotranspiration, and are at the lowest levels in autumn. Hydrographs of wells in Bear Creek Valley exhibit this characteristic seasonal variation.

If any relations could be identified between depth to water or magnitude of fluctuation of water levels and any physical features such as, well depth, land-surface elevation, or geologic formation, then water levels and fluctuation could be estimated for areas having no wells. The information on relations also could be used to select a small number of representative wells to monitor for water-level fluctuation, rather than random selection or continuous monitoring of a large number of wells. Once the magnitude of fluctuation in a well is determined, measurements can be made at longer time intervals. Magnitude of fluctuation in wells was used in this investigation to determine a tolerance for error in head matches during model calibration.

Water levels have been measured at weekly intervals since 1984 for many wells in Bear Creek Valley (Martin Marietta Energy Systems, Inc., written commun., 1987). Hydrographs of 79 of the wells were compared for pattern and amplitude of water-level fluctuations. The hydrographs were separated into three groups: Group A—wells having a very small amplitude and range of fluctuation; Group B—wells having a larger amplitude and range than Group A and showing a pattern of seasonal fluctuation; and Group C—wells having a large amplitude and range of fluctuation, a pattern of seasonal fluctuation, and long, smooth recessions during periods of low rainfall. A representative hydrograph from each group is shown in figure 7 compared to average-monthly (for the period 1956 to 1985) and measured-monthly precipitation. Maximum fluctuation of water levels in these wells for the period of record (fig. 7) is 3 feet (GW-66, Group A), 10 feet (GW-2, Group B), and 18 feet (GW-30, Group C).

Statistical tests (two-tailed t-tests and correlations) were applied to determine whether the mean water level or magnitude of water-level fluctuations for wells in these groups could be related to well depth (depth to the top of the screened interval), depth to water, geologic formation, or elevation of land surface. Standard deviations of the water levels measured in each well were calculated as an indication of variability (or magnitude of the water-level fluctuations). Comparisons were made between the means of standard deviations (Xs) in order to test for differences in variability of water levels between these selected groupings of wells.

The Xs of wells in regolith was compared to the Xs of wells in bedrock using a t-test. The test indicated no significant difference at the 99-percent confidence interval between the variability of water levels in regolith and in bedrock (table 2, Xs of water levels). Differences between Groups A, B, and C could then be tested further without regard to whether the hydrograph represented water levels in regolith or bedrock. The results of t-tests between the groups (A-B, B-C, and A-C) indicate that there are significant differences in the variability of water levels of the three groups (table 2, Xs of water levels). Separation of the hydrographs into the three



Precipitation data from National Oceanic and Atmospheric Administration (1984, 1985, 1986)

Figure 7.--Water levels in observation wells representing Groups A, B, and C, and precipitation at Oak Ridge, Tennessee.

groups, which are based on similarities in water-level fluctuation, was considered valid.

Although there is overlap in the ranges of land-surface elevation of the wells in each group, the mean elevation of wells in Group C, wells having the largest amplitude of fluctuation, is the highest (table 3); in Group B, wells having an intermediate amplitude, mean elevation is intermediate; and in Group A, wells having the smallest amplitude, mean elevation is lowest. Similarly, the mean depth below land surface to water for each group is deepest for Group C, intermediate for Group B, and shallowest for Group A (table 3). T-tests, used to test whether these apparent differences between the groups for land-surface elevations of the wells and mean water levels are significant (table 2), indicate no significant difference in land-surface elevation between Groups A and B, but differences are significant between Groups A and C and between Groups B and C.

In summary, t-tests indicated (1) no significant difference in the mean (water level) of wells in regolith versus those in bedrock, and (2) significant differences in mean water level among Groups A, B, and C. There is also no difference in land-surface elevation between regolith and bedrock wells, probably because the material in which a particular well is set, or depth to which a well is drilled, is not related to a natural condition, but rather, to selection of a water-producing zone by project personnel.

A correlation coefficient was calculated for each group comparing land-surface elevation and standard deviation of mean water levels in a well (table 4). Group C appears to be the only group for which there might be a linear relation between surface elevation of the well and magnitude of water-level fluctuation. Correlation between land-surface elevation and mean depth to water was also tested; other investigators have indicated a relation (Exxon Nuclear Company, Inc., 1978, p. 3.5-8, 3.5-9). The best, although not strong, correlations were for Group B and for the grouping of all wells (table 4). Overall there are slightly better correlations for mean depth to water than for the standard deviations

Table 2.—*Results of two-tailed t-tests at the 99-percent confidence interval for difference in water-level fluctuations between wells in regolith and bedrock and for wells in Groups A, B, and C*

[X = mean of the means of a variable for wells in a group;
Xs = mean of the standard deviations of a variable in the group]

Comparison groups		t-statistic					
		X of water levels	X of land-surface elevations	X of depths to top of screen	Xs of water levels	Critical value	Degrees of freedom
All wells in regolith	All wells in bedrock.	-0.002	-1.50	-6.0	-0.13	2.64	77
Group A	Group B	-2.94	-.19	-1.76	-8.12	2.65	68
Group B	Group C	-2.24	-3.53	-.85	-5.06	2.70	39
Group A	Group C	-3.78	-4.40	-.60	-9.49	2.69	45

Table 3.—*Mean and standard deviation of land-surface elevation, mean depth to water, and mean depth to the top of the screened interval for all wells and for Groups A, B, and C*

[Range, Negative number denotes water level above land surface.]

Group	Number of observations	Land-surface elevation (in feet above sea level)			Mean depth to water (in feet below land surface)			Mean depth to top of screened interval (in feet below land surface)		
		Range	Mean	Standard deviation	Range	Mean	Standard deviation	Range	Mean	Standard deviation
A	38	1,002.8 - 886.8	944.0	29.1	-1.2 - 41.6	7.7	1.03	6.0 - 195.1	41.2	47.9
B	32	1,004.5 - 900.5	945.5	35.3	4.3 - 46.4	14.2	2.03	3.0 - 87.4	25.2	19.7
C	9	1,010.5 - 969.9	988.5	16.6	11.9 - 27.0	20.8	3.87	6.0 - 41.9	31.3	16.7
All wells	79	1,010.5 - 886.8	949.7	33.5	-1.2 - 46.4	14.0	1.76	3.0 - 195.1	33.6	36.5

of the mean water levels, with the exception of Group C. For both comparisons of wells in Group C, the correlations are poor for a relation between depth to water in a well and the land-surface elevation of that well. The poor correlation may be partly due to the small number of wells drilled on the ridges. Correlations to test for a relation between water levels and geologic unit (not shown) were similarly poor, which may indicate a consistent, continuous flow system across the units.

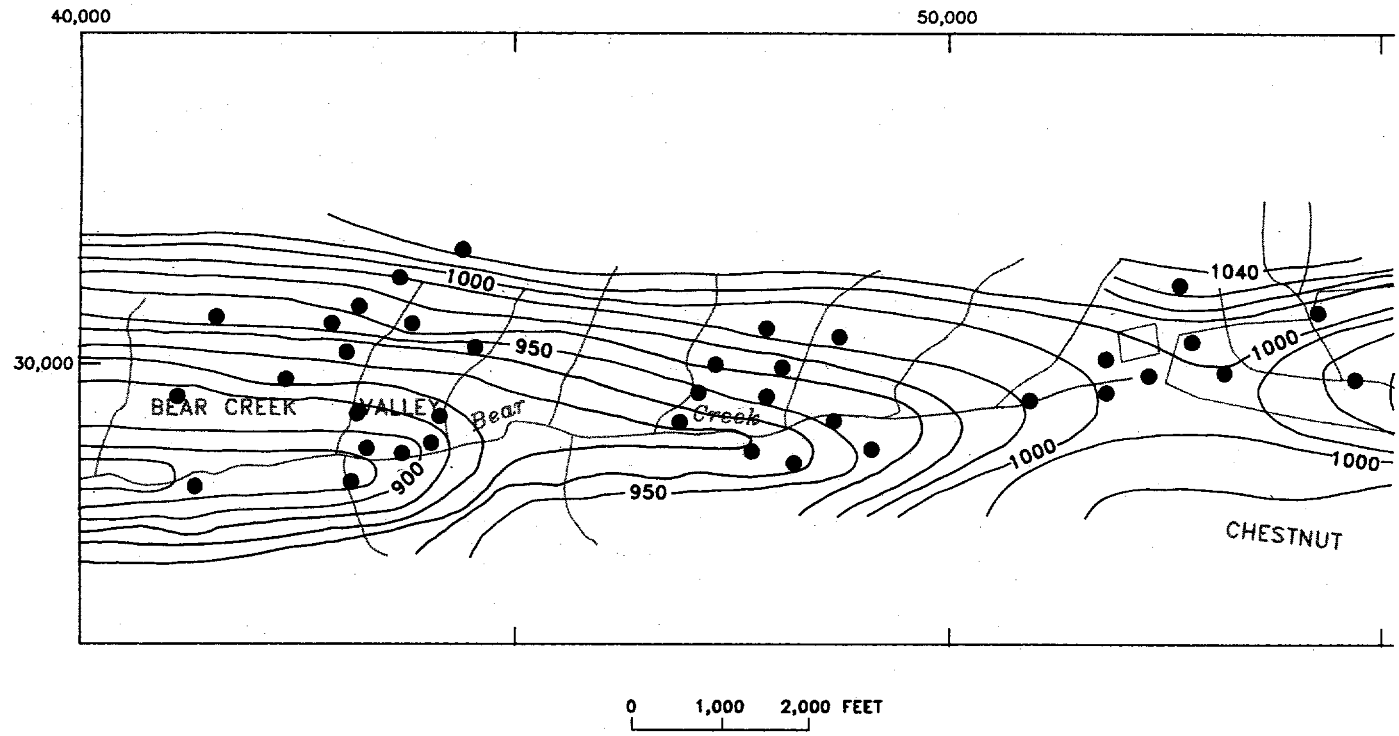
Table 4.—*Correlation coefficients for the relation of land-surface elevation to water levels in all wells and wells in Groups A, B, and C*

Group	Number of observations	Correlation coefficient for land-surface elevation of a well and: Standard deviation from mean water level	Mean depth to water	Mean depth to top of screened interval
A	38	-0.16	0.45	0.28
B	32	-.09	.55	-.14
C	9	.79	.10	-.43
All wells	79	.32	.52	-.09

GROUND-WATER FLOW AND POTENTIAL CONTAMINANT PATHWAYS

Precipitation recharges the shallow ground water, which flows in the regolith and weathered bedrock from the ridges and drainage divides toward the streams as indicated by the water-table configuration (fig. 8). Bear, Grassy, East Fork Poplar, and Scarboro Creeks are the primary drains for shallow ground water in Bear Creek Valley, and a small tributary to the Clinch River is the drain for Union Valley. Water percolating through the regolith recharges the bedrock. Ground-water flow and potential contaminant pathways investigated in this study are on a regional scale and localized behavior of contaminants is beyond the scope of the investigation.

The main ground-water flow component is normal to strike and toward the major streams in the valley. The minor component of shallow flow that corresponds to strike (parallel to the axis of the valley) is controlled by ephemeral streams that cut across strike at regular intervals. Results of the cross-sectional modeling (Bailey, 1988) show that ground water flows to the Maynardville Limestone from both sides of the valley (fig. 9). The Maynardville Limestone is



EXPLANATION

——1000— — WATER-LEVEL CONTOUR--Shows altitude of water table. Dashed where approximately located. Contour intervals 10 and 40 feet. Datum is sea level

B————*B'* TRACE OF GEOLOGIC SECTION (See figure 10.)

● OBSERVATION WELL

Note: Grid coordinate system unique to this map (S-16A)

Figure 8.--Water-table configuration for a segment of Bear Creek Valley, October 1986.

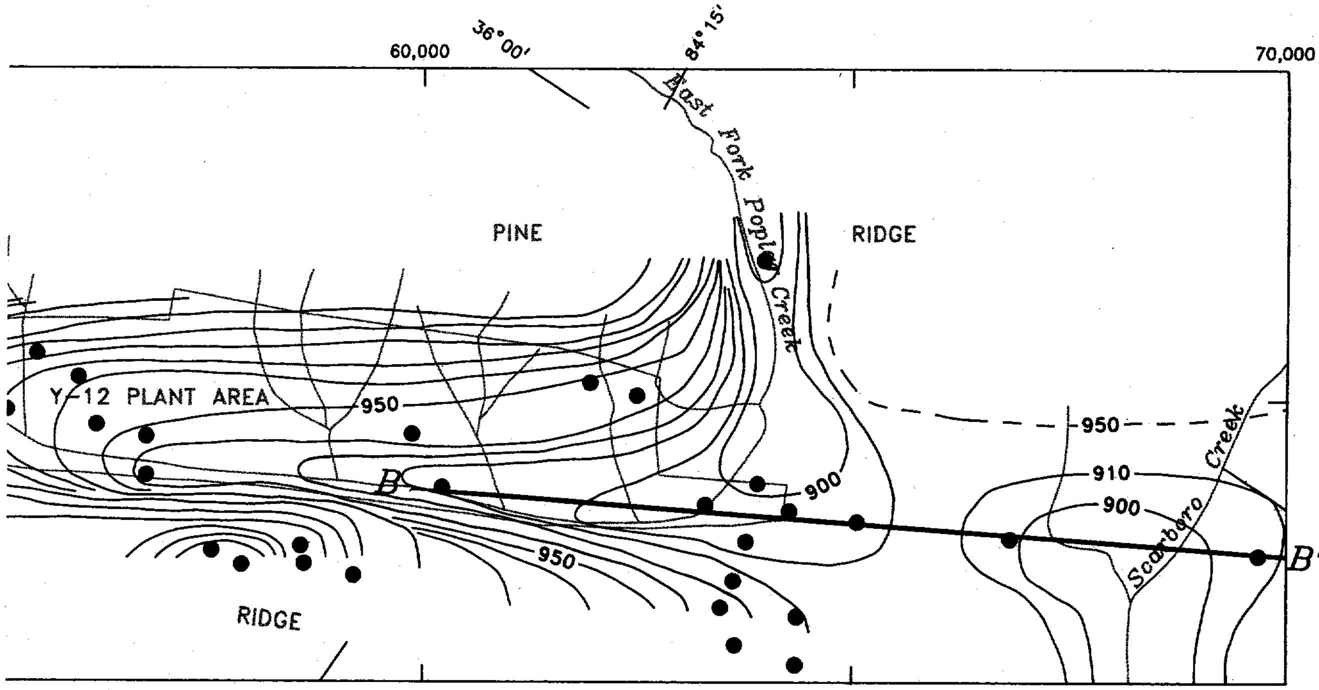


Figure 8.--Water-table configuration for a segment of Bear Creek Valley, October 1986--Continued.

the most significant geologic unit in terms of ability to readily transport contaminants because it contains numerous, large solution channels that are interconnected both along and normal to strike. Contaminants in the ground water could reach the Maynardville and be transported along strike by ground water or be discharged into Bear Creek. Ultimately the main streams, ephemeral streams, and springs are the recipients of most ground-water flow and, therefore, of any contaminants in the ground water.

Significant flow in the valley appears to be approximately limited to the upper 400 feet of geologic materials, although King and Haase (1988, p. 48) estimate the deepest extent of the flow system to be between 500 and 700 feet below land surface. After drilling or pumping, water levels in wells at and below 400 feet in depth recover very slowly, which indicates very low hydraulic conductivity (Geraghty & Miller, Inc., 1987). King and Haase concluded that the distribution of hydraulic conductivities with depth is irregular, and the highest values were measured in structurally disturbed zones in the Maynardville Limestone, Copper Ridge Dolomite, and the uppermost Rome Formation. This conclusion was based on packer testing of intervals ranging in depth from 100 to 1,200 feet in six coreholes across Bear Creek Valley.

Ground-water divides coincide with divides for surface drainage. Scarboro Creek drainage basin appears to be an effective ground-water divide between the East Fork Poplar Creek and Union Valley drainage; therefore, the flow system in Union Valley is separate from and unaffected by flow in Bear Creek Valley. However, deeper

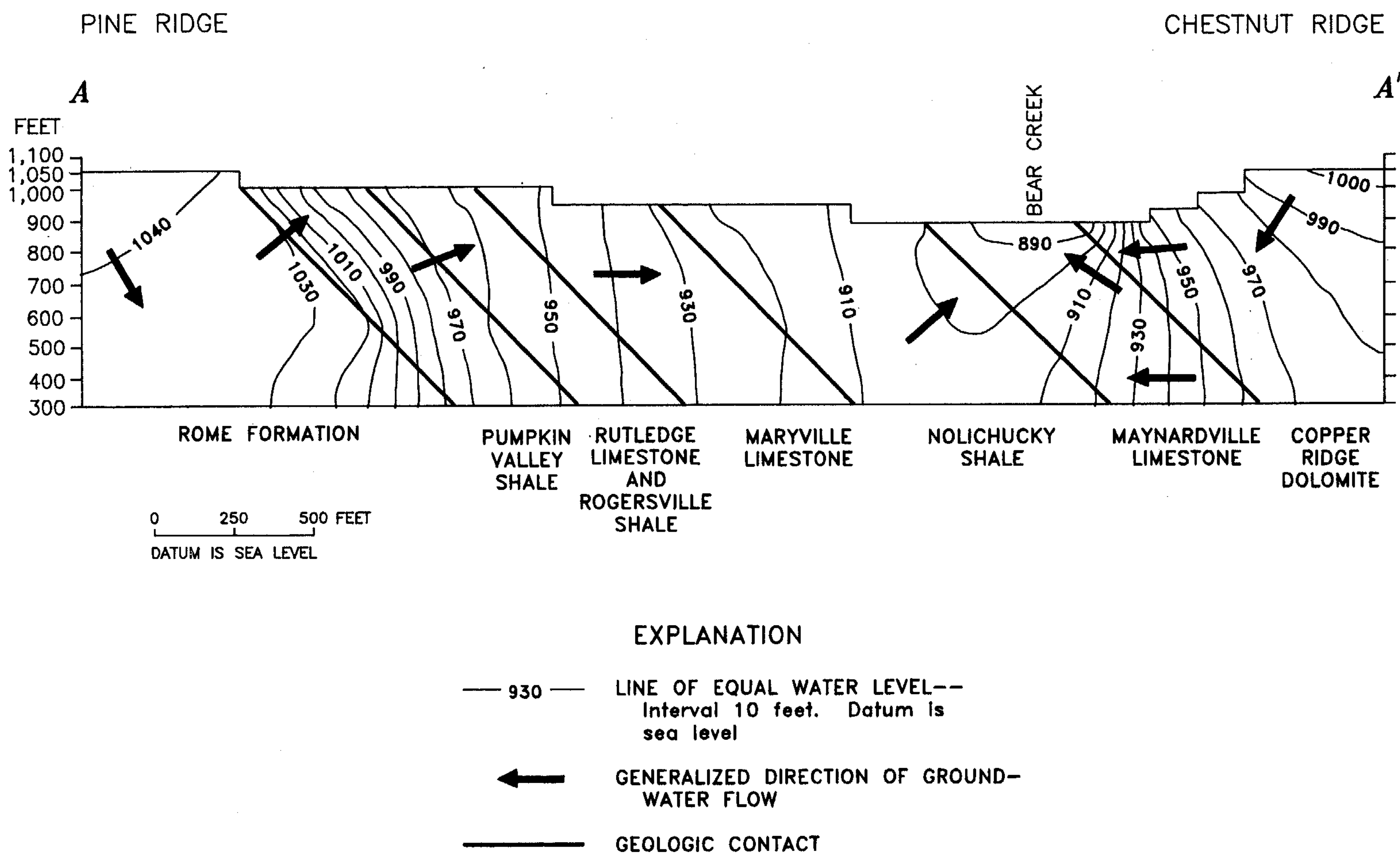


Figure 9.--Model-simulated water levels and direction of ground-water flow along section A-A'.

ground-water flow from the East Fork Poplar Creek basin into Scarboro Creek basin is possible (fig. 10). The strike-parallel hydrogeologic section shows hydraulic potential from the lower Maynardville Limestone and the Nolichucky Shale in the area of New Hope Pond toward Scarboro Creek. This is a two-dimensional perspective of a complex, three-dimensional flow field, and may not represent completely ground-water flow in that area. However, because of the fractured and solutioned nature of the Maynardville Limestone, sources of contaminants in the area, and a potential for flow from the ORR area into Scarboro Creek basin, this area should be considered a possible contaminant pathway.

Ground-water flow over short distances and on a smaller scale than investigated in this study is very complex and is affected by flow cells created by topographic irregularities, small streams, and locally higher hydraulic conductivity within solution, joint, and fracture zones. The contact between the Rome Formation and the Pumpkin Valley Shale is an example of effects of locally high permeability that was not incorporated into the regional flow model of this investigation. Wells in the contact zone are typically flowing wells and heads are higher than the general potentiometric surface (Exxon Nuclear Company, Inc., 1978, p. 3.5-8). King and Haase (1988, p. 39) also report high hydraulic-head data and measured higher hydraulic-conductivity values in the uppermost Rome Formation than at shallower and deeper rock intervals. This more permeable zone is recharged on Pine Ridge and the steep dip of the rocks and low conductivity of the overlying Pumpkin Valley Shale contribute to the artesian flow.

GEOCHEMISTRY OF GROUND WATER

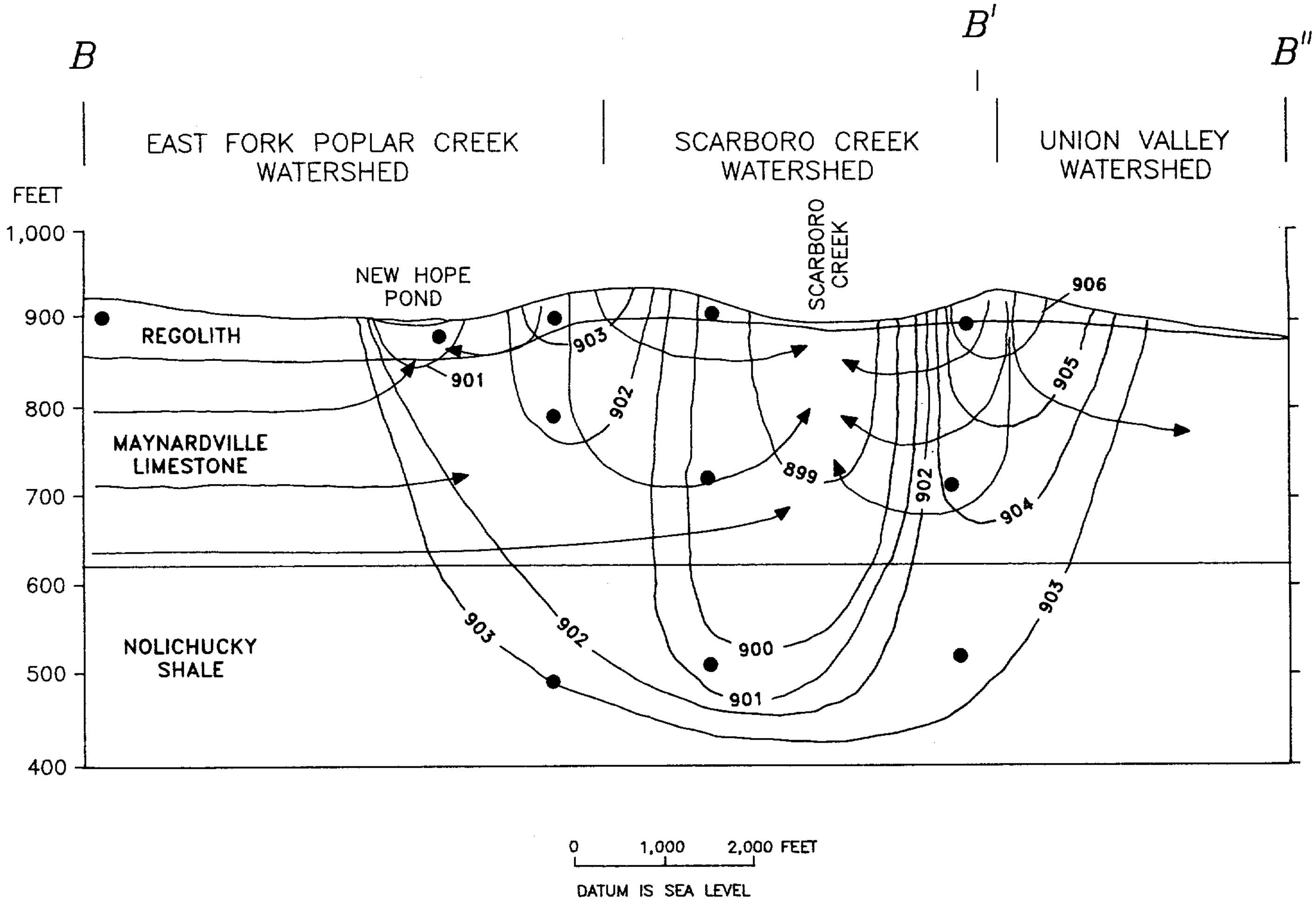
Chemical analyses of water and geochemical interpretations provide additional insight in understanding the ground-water flow system. Chemical data were from wells primarily clustered in the disposal areas where natural water chemistry can be obscured by local contamination. However, the distribution of contaminants can also be used to interpret ground-water flow paths.

General findings of other investigators were that: (1) nitrate concentrations are high in the shallow ground water beneath the S-3 ponds and are detected in water in the bedrock at depths greater than 500 feet below land surface; (2) radionuclides, heavy metals, and volatile organic compounds (VOC's) are major contaminants in both the shallow and deep ground water beneath the Oil Landfarm and the Burial Grounds (Geraghty and Miller, Inc., 1987); and (3) elevated concentrations of chloride in shallow ground water at a disposal site on Chestnut Ridge can be attributed to leaching of road salts used during winter months (Haase and others, 1987a).

DATA BASE

Chemical analyses for ground water from Bear Creek Valley were performed by a variety of private, state, and federal laboratories. More than 500 chemical analyses were available during the period of investigation. Many of the analyses were incomplete, and thus, were of limited use for geochemical interpretations. Of the more than 500 chemical analyses, 142 were used in geochemical interpretations. Analyses were selected if well-inventory data of depth, geologic unit, and location were known. The number of analyses selected was further refined by the availability of pertinent chemical data. Only analyses with pH values greater than 4.0 but less than 11.0 were selected.

For wells having more than one analysis, the most complete or most recent analysis was selected (*Appendix A*). The chemical data were collected from August 1985 to May 1987.



EXPLANATION

—903— LINE OF EQUAL POTENTIAL--Interval 1 foot. Datum is sea level

DIRECTION OF FLOW

GEOLOGIC CONTACT

ELEVATION OF SCREENED OR OPEN INTERVAL OF WELL

Figure 10.--Direction of ground-water flow in the Maynardville Limestone and Nolichucky Shale along section B-B'-B'', September 1986.

GEOCHEMICAL METHODS

Piper diagrams, maps of the distribution of chemical constituents in ground water, and geochemical models were used to represent both the hydrogeology and chemical evolution of ground water. Piper diagrams provide a graphical aid in determining differences in water chemistry and the most likely causes of those differences in any hydrologic system. Mapping and contouring of the chemical data show the areal distribution of constituents, which allows trends and gradients of these constituents to be related to the geology and ground-water flow system. Dissolved solids and dissolved calcium were the most commonly measured constituents in all wells, and provided the best areal distribution for mapping. Geochemical models were used to test hypotheses regarding specific chemical reactions that influence the chemical evolution of ground water.

CHEMICALLY DISTINCT ZONES

At least two chemically distinct water-bearing zones were identified in the steeply-dipping sandstone, limestone, and shale in Bear Creek Valley (fig. 11) on the basis of dissolved-solids concentrations. One zone is less than 50 feet deep, and the other is 50 to 500 feet deep. The zone less than 50 feet deep consists largely of regolith and shallow, weathered bedrock (and fill material in the Y-12 Plant area), and contains water that is chemically distinct from water from deeper wells. Analyses of water samples from wells deeper than 50 feet were used to characterize the geochemistry of a water-bearing zone between 50 and 500 feet below land surface. Areal geochemical data were insufficient at varying depth intervals to distinguish chemically different water-bearing zones within the 50- to 500-foot deep zone. Chemical data of water from zones deeper than 500 feet are not discussed because data from only two wells were available.

Rothschild and others (1984) reported statistical correlations between geologic formations and ratios of magnesium and calcium concentrations, and of silica concentration in ground water for 43 wells in 3 formations of the Conasauga Group, which suggests lithologic control on ground-water chemistry. Ground water may develop chemical signatures of the rock chemistry as the water passes through rocks of different composition. However, the correlations could also result from varying degrees of chemical evolution manifest in differences in the chemistry of shallow and deep water-bearing zones.

SHALLOW WATER-BEARING ZONE

In water from the shallow zone, concentrations of dissolved solids range from 100 to 500 mg/L over most of the area. However, concentrations of dissolved solids are greater than 1,000 mg/L in water from several of the wells, and at a few wells, concentrations are greater than 10,000 mg/L (*Appendix A*). The large increases in dissolved constituents are principally in calcium and chloride, nitrate, or sulfate (fig. 12). Although calcium data are not available for every analysis, elevated calcium concentrations suggest that localized, surficial effects influence the chemistry of shallow ground water. Recharge water contains concentrations of dissolved solids of less than 100 mg/L and is exemplified by water from Bear Creek (fig. 13), which is dominated by calcium and bicarbonate (Pulliam, 1985b). Two primary chemical processes cause shifts in the location of data points on the quadrilinear part of the Piper diagram (fig. 12) outside the area defined by recharge water chemistry. Natural chemical evolution (as water moves through the rock and dissolves minerals, usually dolomite and some gypsum or anhydrite) is evident in a few of the samples (fig. 12) from wells shallower than 50 feet deep. Salting of roadways using $CaCl_2$ "salt" to melt snow in Bear Creek Valley (Haase and others, 1987a) and migration of acidic nitrate wastes

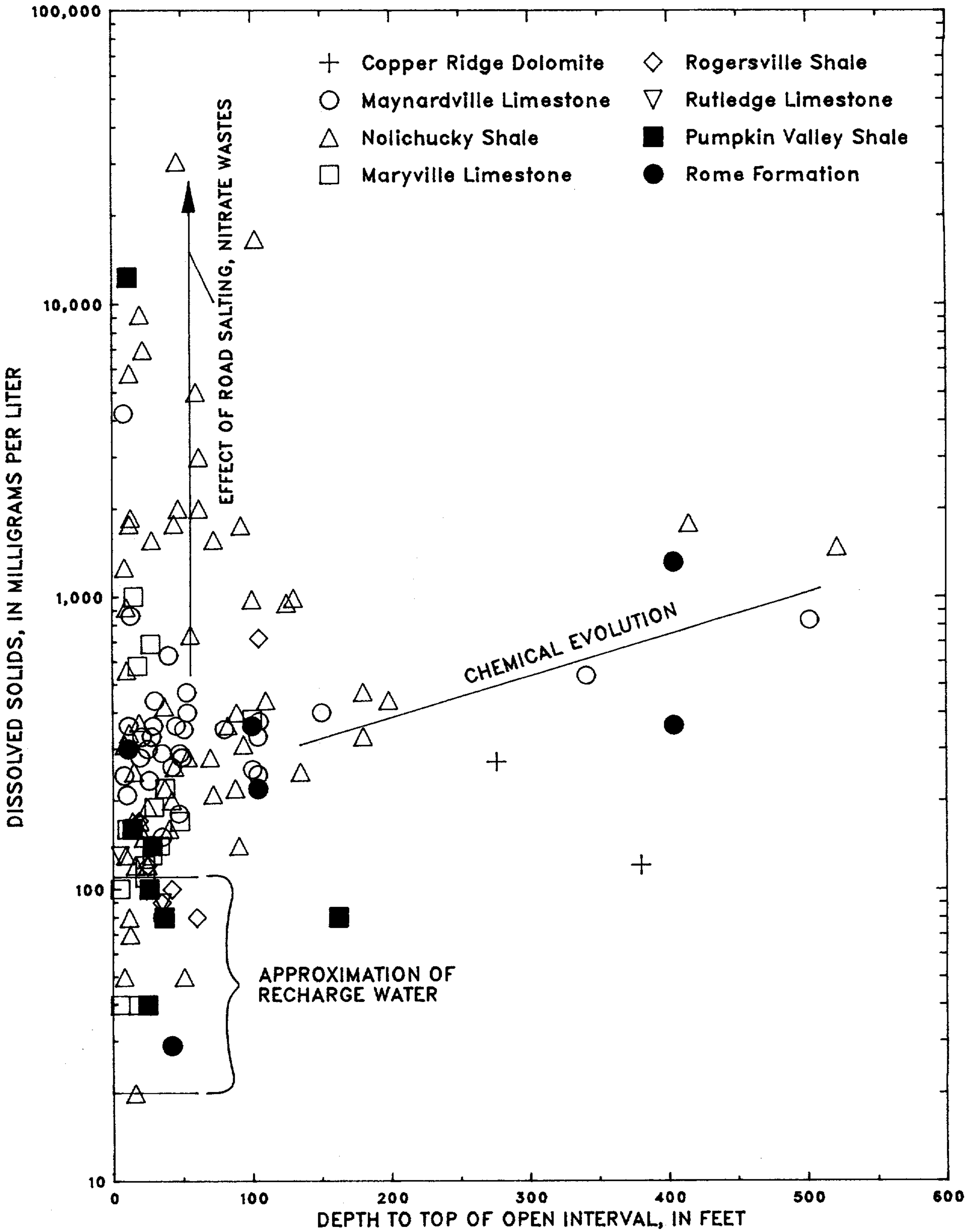


Figure 11.--Relation between concentration of dissolved solids and depth of ground water.

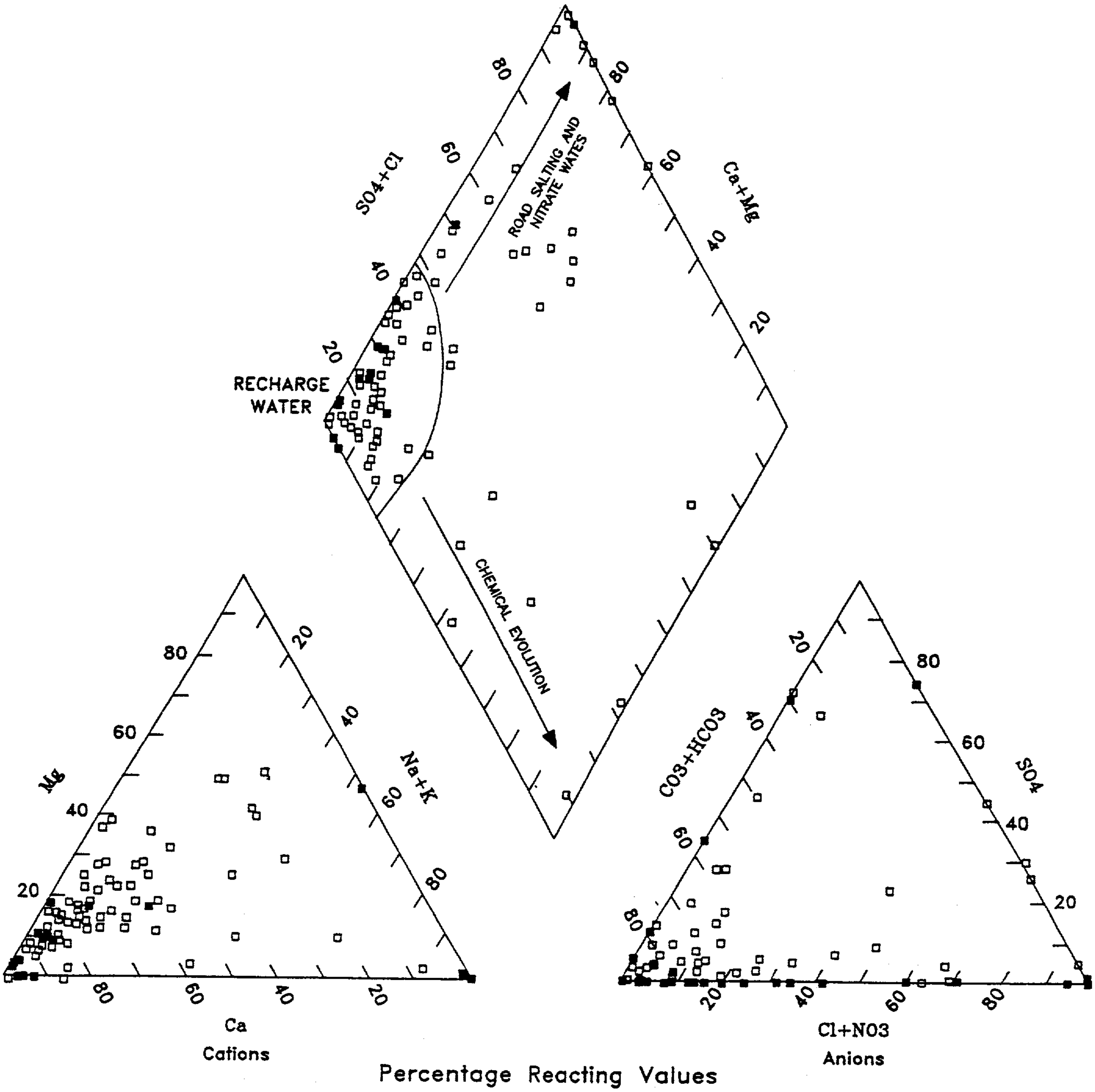


Figure 12.--Chemical composition of water from wells shallower than 50 feet.

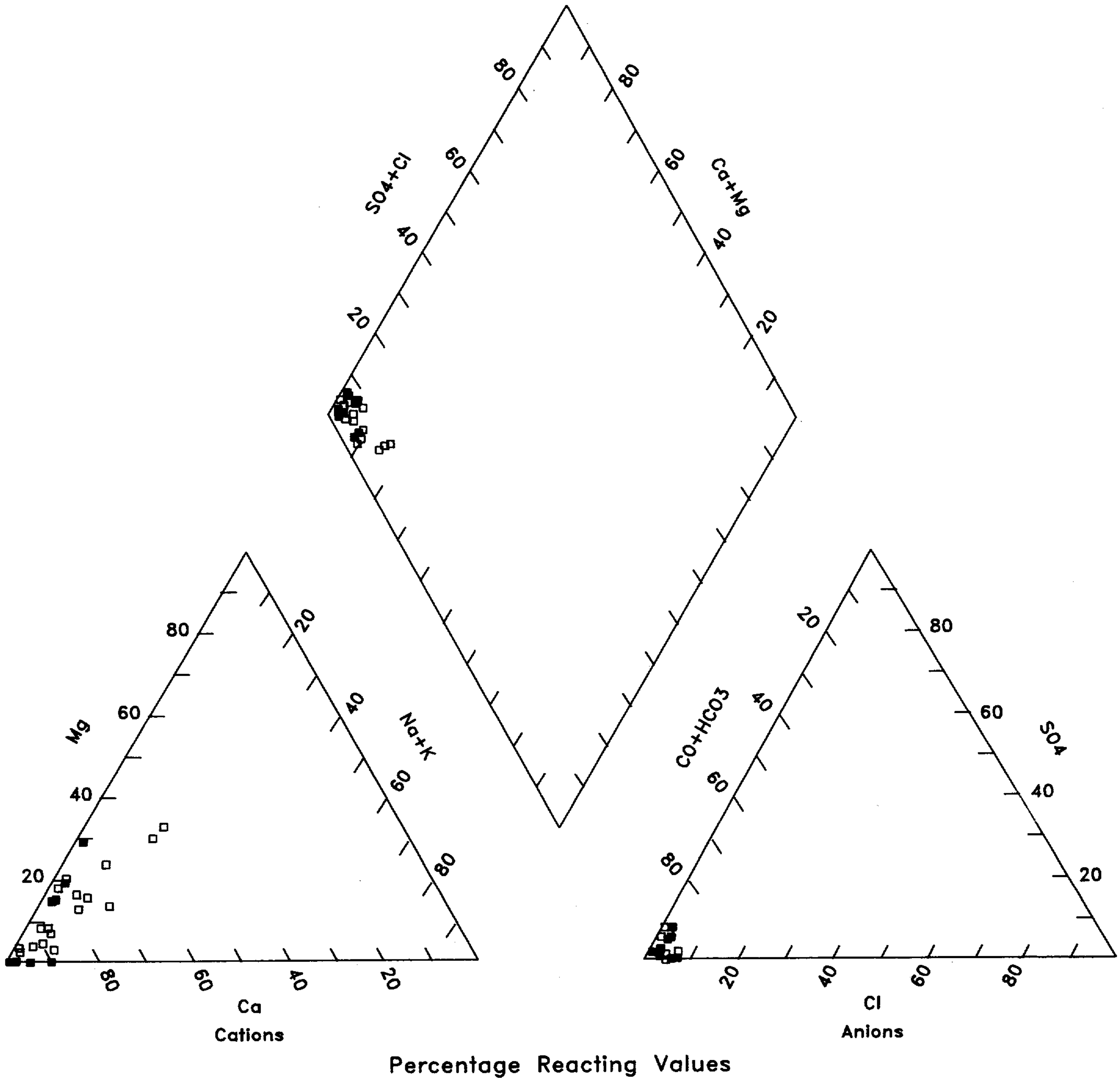


Figure 13.--Chemical composition of water from Bear Creek, September 1984. (Data from Pulliam, 1985b.)

from S-3 ponds (Geraghty & Miller, Inc., 1985a, p. 8-15 through 8-19) are man-induced activities that increase the solute concentrations in ground water.

Principal data coverage of concentrations of dissolved solids for wells less than 50 feet deep is for three areas southwest of the Y-12 Plant (fig. 14). The area northeast of the S-3 ponds contains elevated concentrations of dissolved solids (greater than 10,000 mg/L) in the shallow water-bearing zone. A sharp change in concentration of dissolved solids in shallow ground water in the Oil Landfarm area indicates localized influences of salt loading or waste leachates mixed with some natural chemical evolution of ground water. In the Burial Grounds, solute loading or leachate migration has produced an area of high concentration of dissolved solids (greater than 900 mg/L). Adjacent to that area is an area of lower concentrations that may indicate localized recharge to shallow ground water.

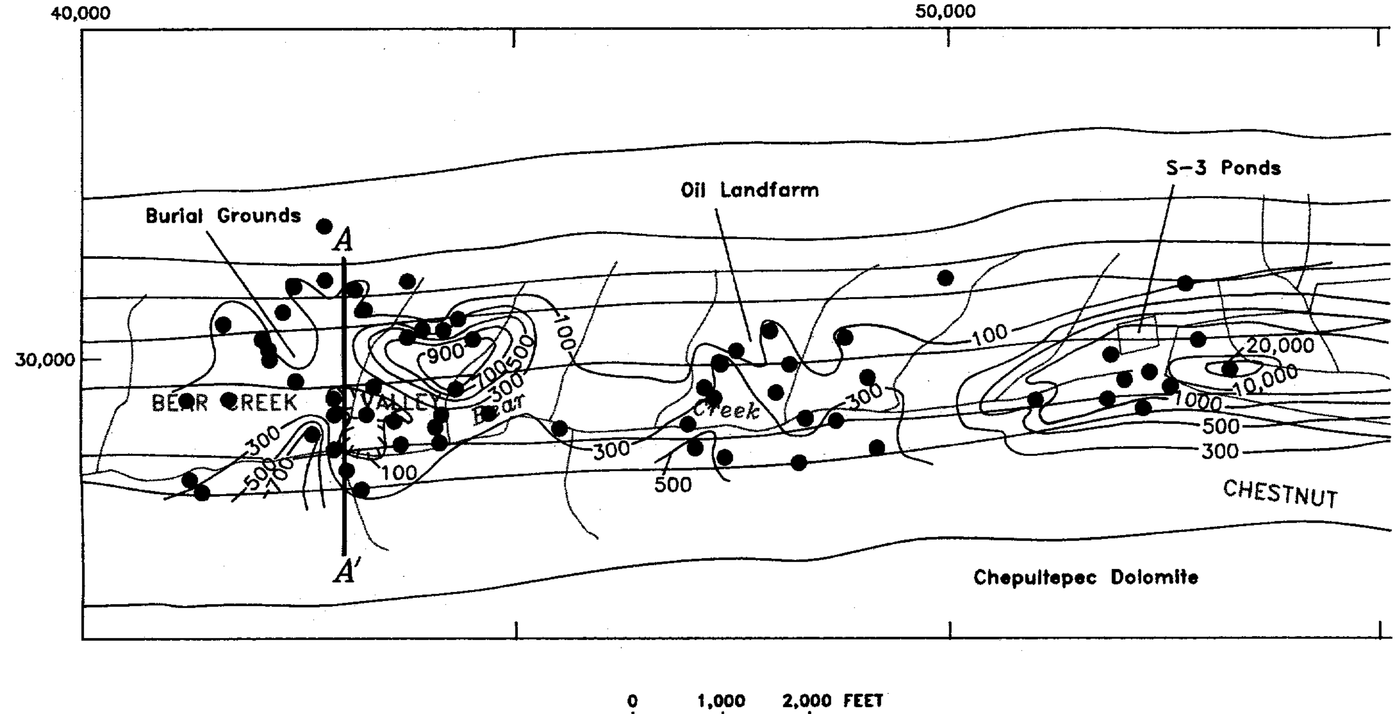
Dissolved-calcium concentration in water from wells less than 50 feet deep has a distribution similar to dissolved-solids concentrations (fig. 15). Recharge water is low in calcium but the concentration increases as calcite and dolomite dissolve. An area of high calcium concentration (as well as high concentration of dissolved solids) is in the Y-12 Plant area northeast of the S-3 ponds. Dissolved-calcium concentration increases away from the dissolved-solids low in the Burial Grounds described previously. This increase supports the concept of localized recharge. Elevated calcium concentrations in the Y-12 Plant area (greater than 10,000 mg/L), and the Burial Grounds (greater than 200 mg/L) supports the interpretation of apparent salt-loading of the shallow water-bearing zone as demonstrated by elevated dissolved-solids concentrations at these same locations.

DEEP WATER-BEARING ZONE

Concentrations of dissolved solids in water from wells deeper than 50 feet may indicate localized surficial influences (fig. 11) from road salt or buried wastes, but generally reflect chemical evolution along deeper flowpaths. The results of natural evolution of ground water would be expected in all of the formations, but is prominent in samples from deep wells in the Rome Formation, Nolichucky Shale, Maynardville Limestone, and Copper Ridge Dolomite (fig. 11). Natural chemical evolution is evident where trends in water chemistry are toward sodium and bicarbonate dominance, and sometimes sodium chloride-sulfate dominance (fig. 16). In contrast, recharge areas in the shallow water-bearing zone are represented by calcium bicarbonate dominated chemistry and dissolved-solids concentrations less than about 100 mg/L. A few of the deep wells sampled contain high concentrations of chloride and nitrate anions, which is characteristic of contamination.

Natural chemical evolution is indicated in the area south of the Burial Grounds (fig. 17). Concentrations of dissolved solids in ground water are low near the base of Pine Ridge, indicating recharge, and concentrations increase toward Bear Creek. Concentrations of dissolved solids in wells deeper than 50 feet are elevated within the Burial Grounds (greater than 1,500 mg/L) and in the Y-12 Plant area (greater than 15,000 mg/L), in the same approximate locations as in the shallow water-bearing zone, which indicates that the contaminants in the shallow water-bearing zone are also present in the deep zone in these areas.

From the limited data, concentrations of dissolved calcium in ground water appear to decrease from the ridges toward Bear Creek (fig. 18). This decrease suggests geochemical processes in which calcium is dissolved from minerals during recharge, and then is removed, perhaps by chemical precipitation as calcite, as



EXPLANATION

—— 1000 —— LINE OF EQUAL CONCENTRATION OF DISSOLVED SOLIDS--Hachures indicate depression. Interval, in milligrams per liter, is variable

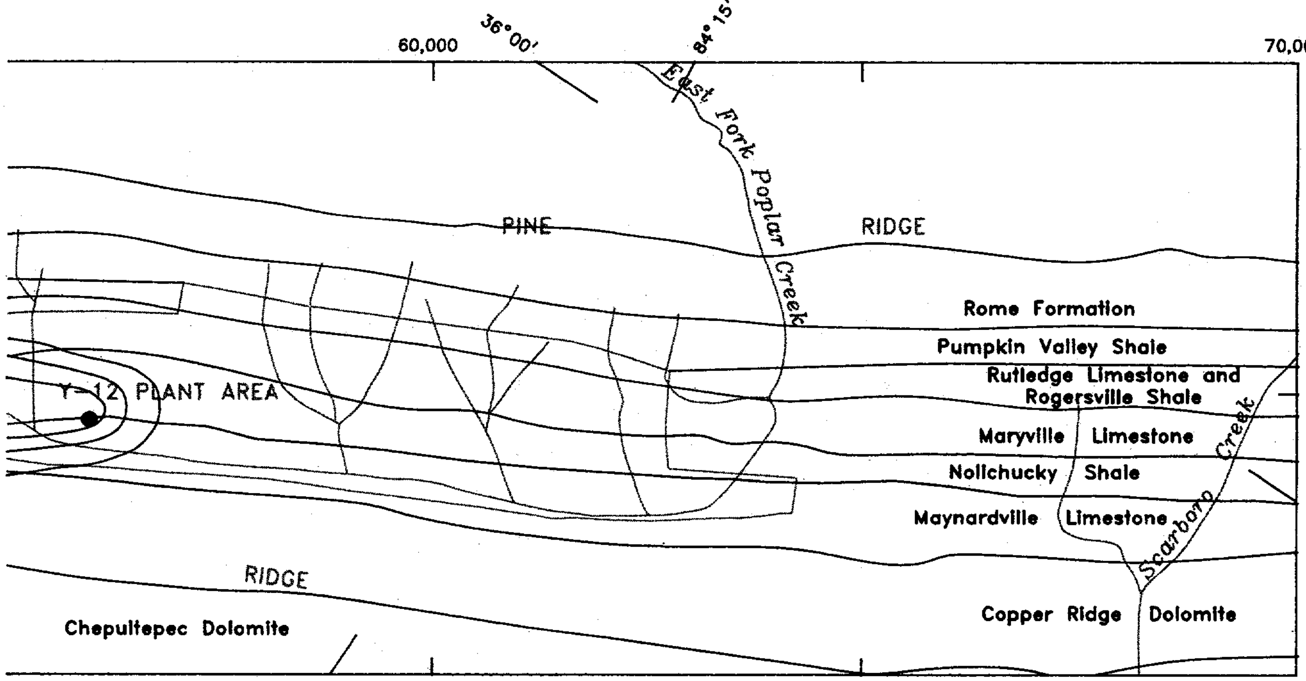
A ——— A′ TRACE OF GEOLOGIC SECTION

——— GEOLOGIC CONTACT

● OBSERVATION WELL

Note: Grid coordinate system unique to this map (S-16A)

Figure 14.--Concentration of dissolved solids in water from wells less than 50 feet deep.



Geology based in part on interpretations by Exxon Nuclear Company, Inc., 1978; R.H. Ketelle and D.D. Huff, 1984; Low Engineering and Testing Company, written commun.,1983; W.M. McMaster, 1962; and G.D. Swingle and E.T Luther, 1964

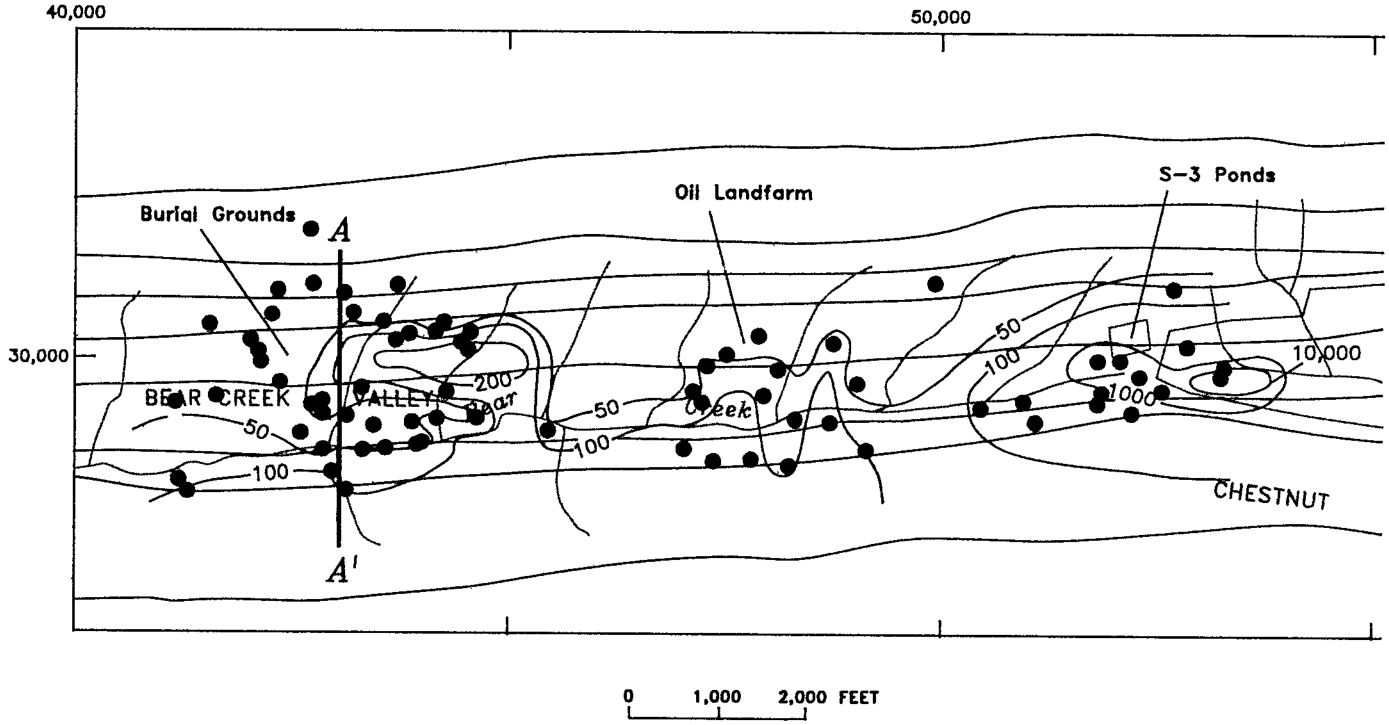
Figure 14.--Concentration of dissolved solids in water from wells less than 50 feet deep--Continued.

ground water flows toward Bear Creek. High concentrations of dissolved calcium in the Burial Grounds (greater than 400 mg/L) and in the Y-12 Plant area (greater than 3,000 mg/L) probably result from contaminant sources previously discussed.

A reversal of the chemical gradient of dissolved solids in the deeper water-bearing zone (fig. 19) suggests upward flow and discharge of ground water from the deep water-bearing zone to the shallow system near Bear Creek, and movement of freshwater from Chestnut Ridge.

GEOCHEMICAL EVOLUTION OF GROUND WATER

Complete chemical analyses collected in April 1987 (*Appendix A*) from selected USGS wells were used in geochemical models to assess data reliability and to test various hydrologic and



EXPLANATION

—— 1000 —— LINE OF EQUAL CONCENTRATION OF DISSOLVED CALCIUM--Dashed where approximately located. Interval, in milligrams per liter, is variable

A————A' TRACE OF GEOLOGIC SECTION (See figure 10.)

● OBSERVATION WELL

Note: Grid coordinate system unique to this map (S-16)

Figure 15.--Concentration of dissolved calcium in water from wells less than 50 feet deep.

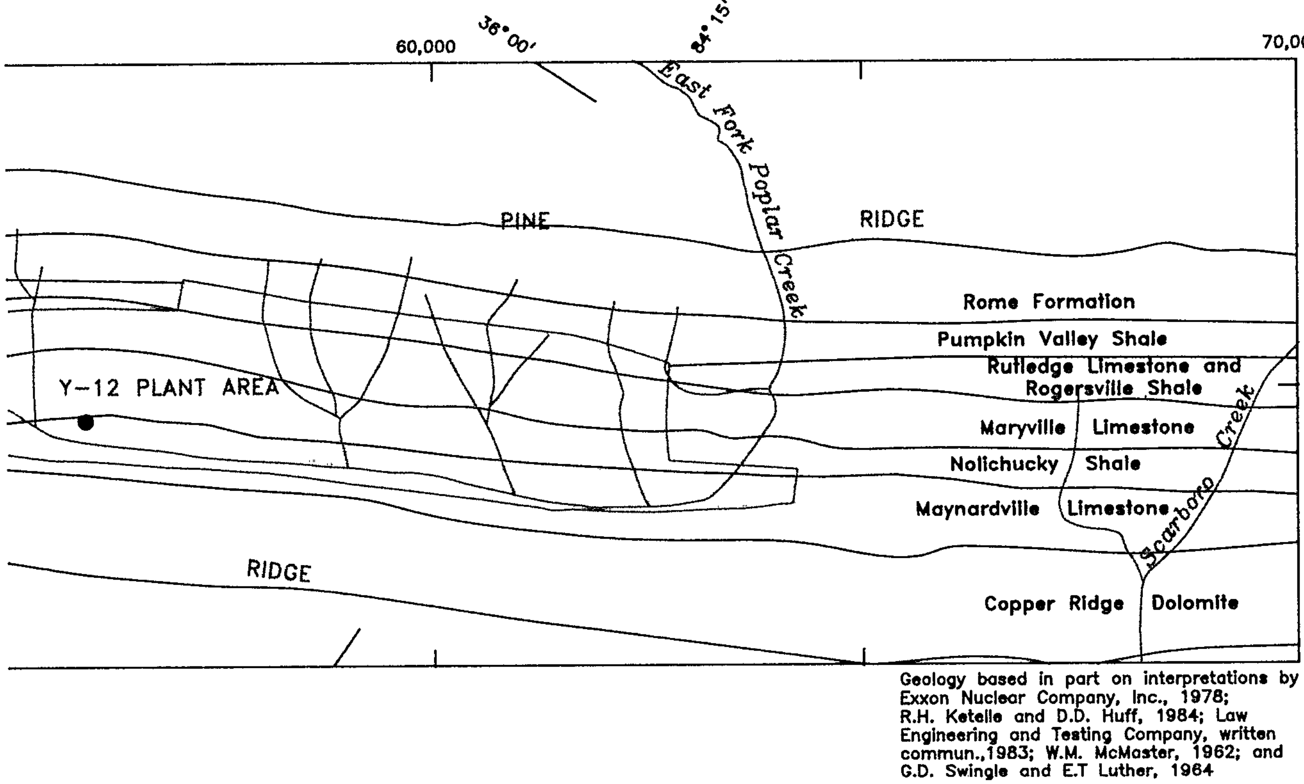


Figure 15.--Concentration of dissolved calcium in water from wells less than 50 feet deep--Continued.

geochemical hypotheses. In addition to the chemical analyses, field pH and bicarbonate concentration of each water sample were measured. Saturation states with respect to possible minerals in the geologic units were determined for samples from each well using the computer code WATEQF (Plummer and others, 1976). Models of chemical evolution of ground water from the Rome Formation, Maynardville Limestone, and Copper Ridge Dolomite were constructed using PHREEQE computer code (Parkhurst and others, 1980). A more complete description of applications of mass transfer models is found in Plummer and others (1983).

The results of the WATEQF calculations indicate that most of the samples of ground water were near saturation or above saturation for the minerals calcite, dolomite, chalcedony, and barite (fig. 20). Saturation of a mineral phase is presumed where the saturation index (SI) is plus or minus 0.1. Gypsum (and anhydrite) were undersaturated in water from all wells, although water from GW-211 (from the Rome Formation) was slightly below saturation. Water from well GW-209, which is in the recharge area of the Rome Formation, was undersaturated with the principal minerals except chalcedony, which indicates that this water has not been in contact

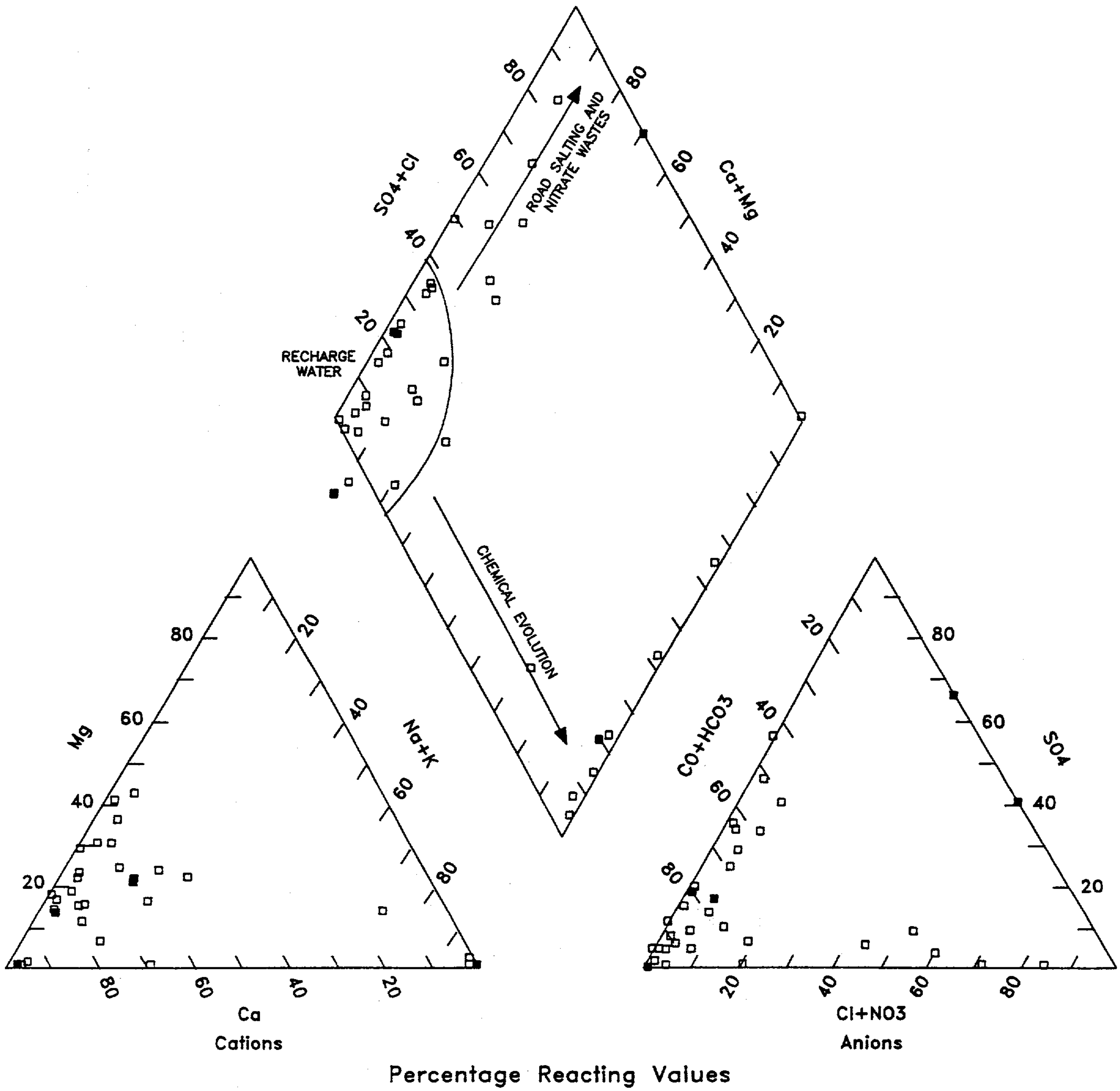


Figure 16.--Chemical composition of water from wells deeper than 50 feet.

with minerals in the rock long enough to reach equilibrium. This particular water chemistry (GW-209) was used to represent recharge water for the mass transfer models of the Rome Formation and the Maynardville Limestone. Wells GW-210 in the Rome Formation, GW-239 and GW-214 in the Nolichucky Shale, and GW-238 in the Maynardville Limestone, were over-saturated with respect to calcite and dolomite compared to other samples, and pH values were basic, ranging from 8.1 to 10.0 units. Although some natural waters may achieve pH values such as these, the saturation states with respect to the carbonate minerals suggest that these measurements are artifacts of well completion and hydroxy salts have been added to the ground water in the vicinity of the well from cement grout. The problem is enhanced by the low production capacity of these wells, less than a half gallon per minute from wells in the Nolichucky Shale; complete purging of the chemical effects of the grout from the well bore and adjacent rock is difficult if not impossible to achieve.

Mass transfer models were developed to simulate chemical evolution of ground water and to test various hypotheses of hydrology and mineral and water interactions. Much additional information, such as complete chemical and mineralogic data from the solid phases of the various aquifers and isotope geochemistry for aqueous and solid phases, is needed to fully support model assumptions. However, some useful interpretations are possible even with these limited modeling efforts.

Mass transfer models based on aqueous equilibrium thermodynamics were constructed using mineral-saturation states of water samples (from WATEQF) as guides, and limited mineralogic evidence and descriptions from the literature for the Rome Formation, Maynardville Limestone, and Copper Ridge Dolomite. Typical chemical reactions modeled are shown in table 5. Temperature was increased in equal increments from 14 $^{\circ}C$ to 16 $^{\circ}C$ in order to simulate the range of ground-water temperatures for the samples.

ROME FORMATION

The chemistry of water from well GW-209 was used to represent the chemistry of water recharging the Rome Formation. In the model, sufficient mineral phases were dissolved to bring the water chemistry to equilibrium with calcite, dolomite, and chalcedony. Chemical evolution was continued to calcite saturation; calcite and dolomite were dissolved as carbon dioxide was added to the system (fig. 21a), presumably from degradation of natural organic matter. This step was necessary to account for increases in dissolved minerals and dissolved inorganic carbon (fig. 21b) in this part of the flow system. The amount of CO_2 is constrained by the amount of dissolved minerals and pH of the water samples. Gypsum (or anhydrite) ($CaSO_4$) was added to simulate observed increases in dissolved sulfate. Magnesium measured in water samples can be accounted for by using a cation exchange reaction with calcium. Significant losses of CO_2, modeled in step 2, were apparent in deeper wells in the Rome Formation. This step produced significant

Table 5.—*Probable chemical reactions in ground water*

DISSOLUTION:

$CO_2(g)(soil) + H_2O = H_2CO_3$

$CaCO_3 + CO_2 + H_2 = Ca^{2+} + 2HCO_3-$
(calcite)

$CaMg(CO_3)^2 + 2CO_2 + 2H_2O = Ca^{2+} + Mg^{2+} + 4HCO_3-$
(dolomite)

$CaSO_4 = Ca^{2+} + SO_42-$
(gypsum)

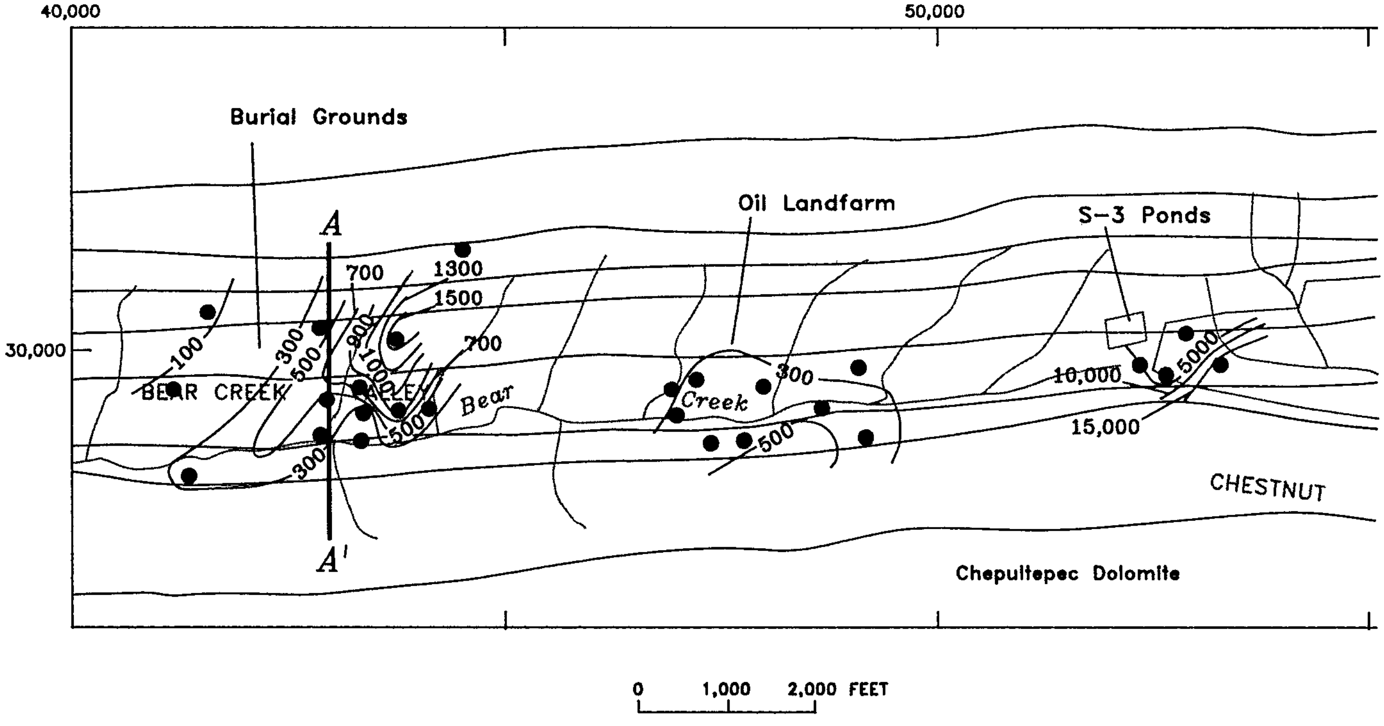
$NaCl = Na^+ + Cl^-$

CATION EXCHANGE:

$Ca^{2+} + Na_2\text{-clay} = 2Na^+ + Ca\text{-clay}$

OXIDATION AND REDUCTION:

$CH_2O + O_2 = CO_2 + H_2O$
(lignite)



EXPLANATION

——1000—— LINE OF EQUAL CONCENTRATION OF DISSOLVED SOLIDS-- Interval, in milligrams per liter, is variable

A————A' TRACE OF GEOLOGIC SECTION

———— GEOLOGIC CONTACT

● OBSERVATION WELL

Note: Grid coordinate system unique to this map (S–16A)

Figure 17.--Concentration of dissolved solids in water from wells deeper than 50 feet.

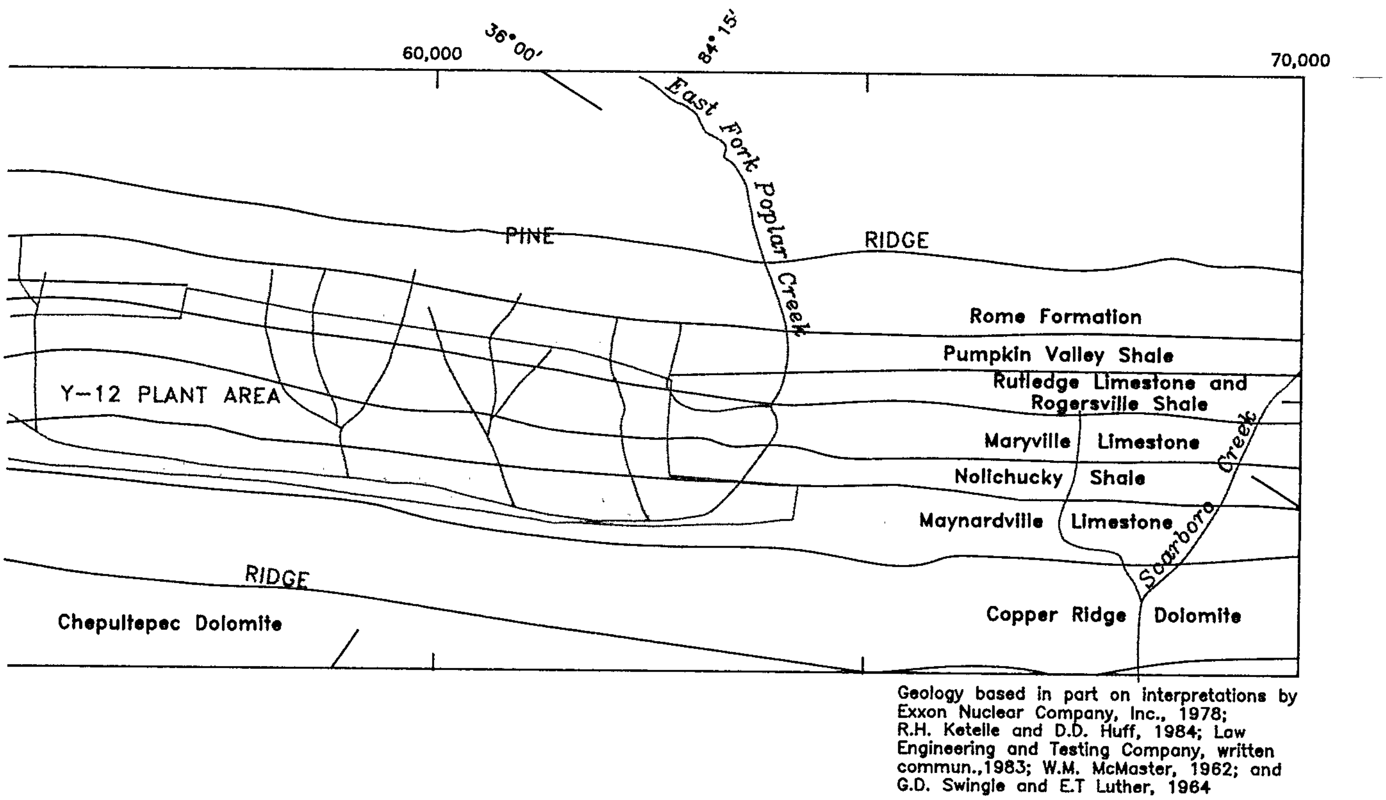
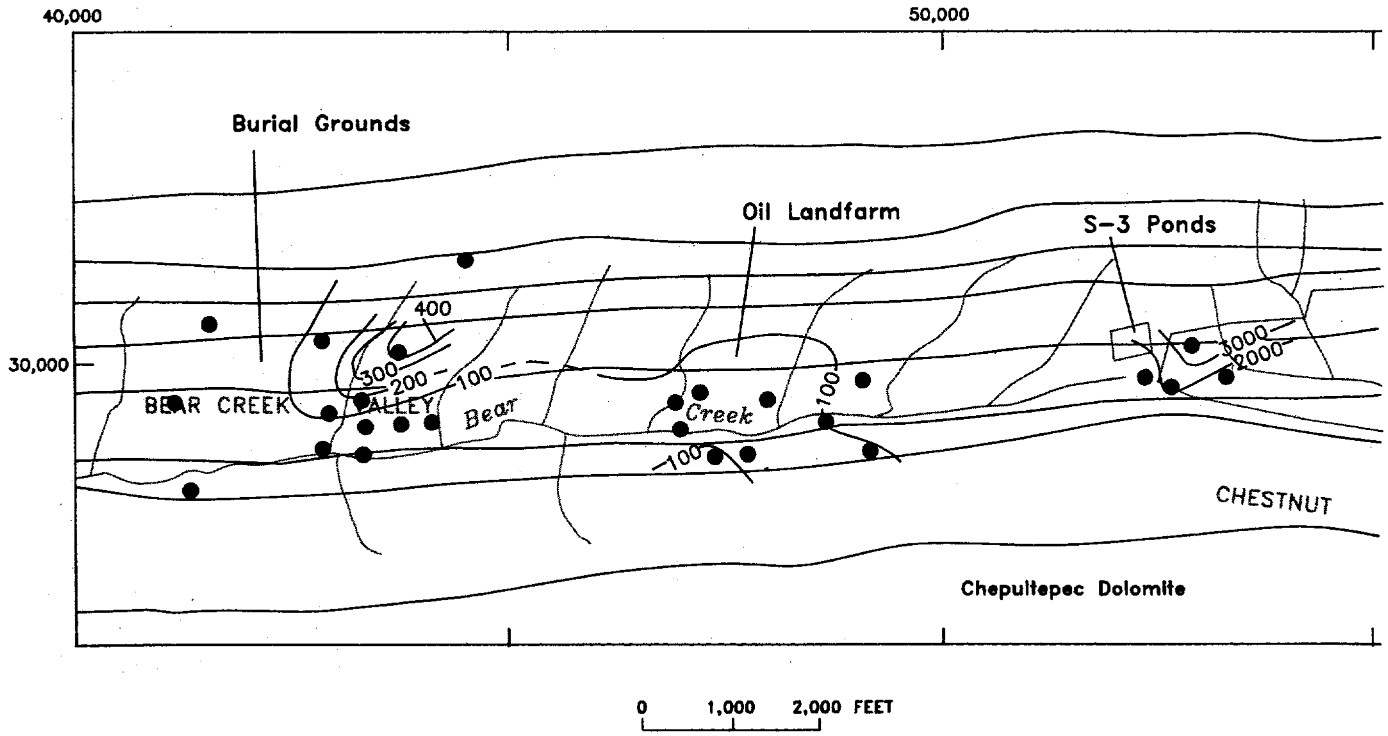


Figure 17.--Concentration of dissolved solids in water from wells deeper than 50 feet--Continued.

precipitation of calcite in the model (*Appendix B*). Because such large losses of CO_2 do not usually occur in ground-water systems that are isolated from the atmosphere, the loss could be explained as gas evasion in the well bore during pumping of a low-yield well. Thus, calcite may not be precipitating in the deeper ground-water system. It is also possible that some mixing with chemically dissimilar solutions from adjacent water-bearing zones may have occurred, producing the observed chemistry of water from the Rome Formation. Determination of the proper chemical model for the Rome Formation would require further study of cycling of stable carbon and sulfur isotopes in the mineral-water system, and solid phase chemistry and mineralogy. Present modeling is preliminary, and should be considered cautiously. Model results were a satisfactory fit to the data (fig. 21).

MAYNARDVILLE LIMESTONE

Water chemistry from well GW-209 served as the recharge water chemistry to the Maynardville Limestone for the model. Equilibrium with calcite, dolomite, and chalcedony were initially established by dissolving calcite and dolomite up to their respective phase boundaries. Subsequently, the model dissolved dolomite and precipitated calcite as CO_2 was added to the system downgradient (*Appendix B*). Some cation exchange, although minimal, was included. In the deeper, and presumably more chemically evolved



EXPLANATION

—— 2000 —— LINE OF EQUAL CONCENTRATION OF DISSOLVED SOLIDS--Dashed where approximately located. Interval, in milligrams per liter, is variable

———— GEOLOGIC CONTACT

● OBSERVATION WELL

Note: Grid coordinate system unique to this map (S–16A)

Figure 18.--Concentration of dissolved calcium in water from wells deeper than 50 feet.

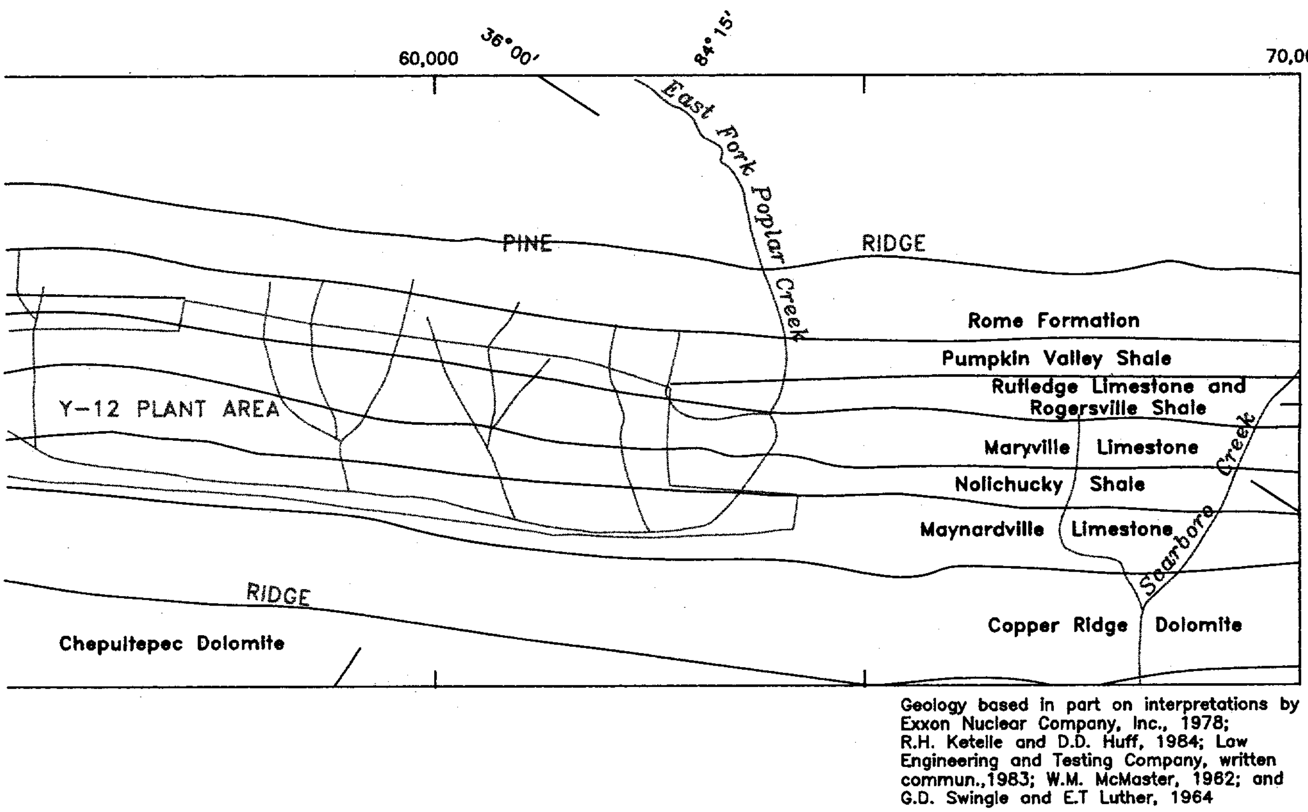


Figure 18.--Concentration of dissolved calcium in water from wells deeper than 50 feet--Continued.

water, sodium chloride and gypsum were added in the second reaction step to more closely simulate the observed chemical data. The reaction path was more straightforward than in the model for the Rome Formation, because no CO_2 decreases were observed in deep wells, although mixing with dissimilar waters is a minor possibility. Some high chloride concentrations in GW-172 and GW-230 indicate that mixing with water from the Copper Ridge Dolomite is more likely than mixing with water from the Rome Formation. Model results were a good fit to observed data from the Maynardville Limestone (fig. 22).

COPPER RIDGE DOLOMITE

Because chemical data from water in the Copper Ridge Dolomite were limited to two partial analyses from wells GW-165 and GW-166 (fig. 3), a very simplified mass transfer model was constructed. Chemical reactions were simulated in a single step (*Appendix B*) from recharge water consisting of pure water plus CO_2 (1.82 moles X 10^{-3}). Results indicate that uptake of SiO_2 to the chalcedony phase boundary is plausible with relatively low levels of dolomite dissolution. Carbon dioxide dissolution, presumably from decay of organic matter, is comparable to

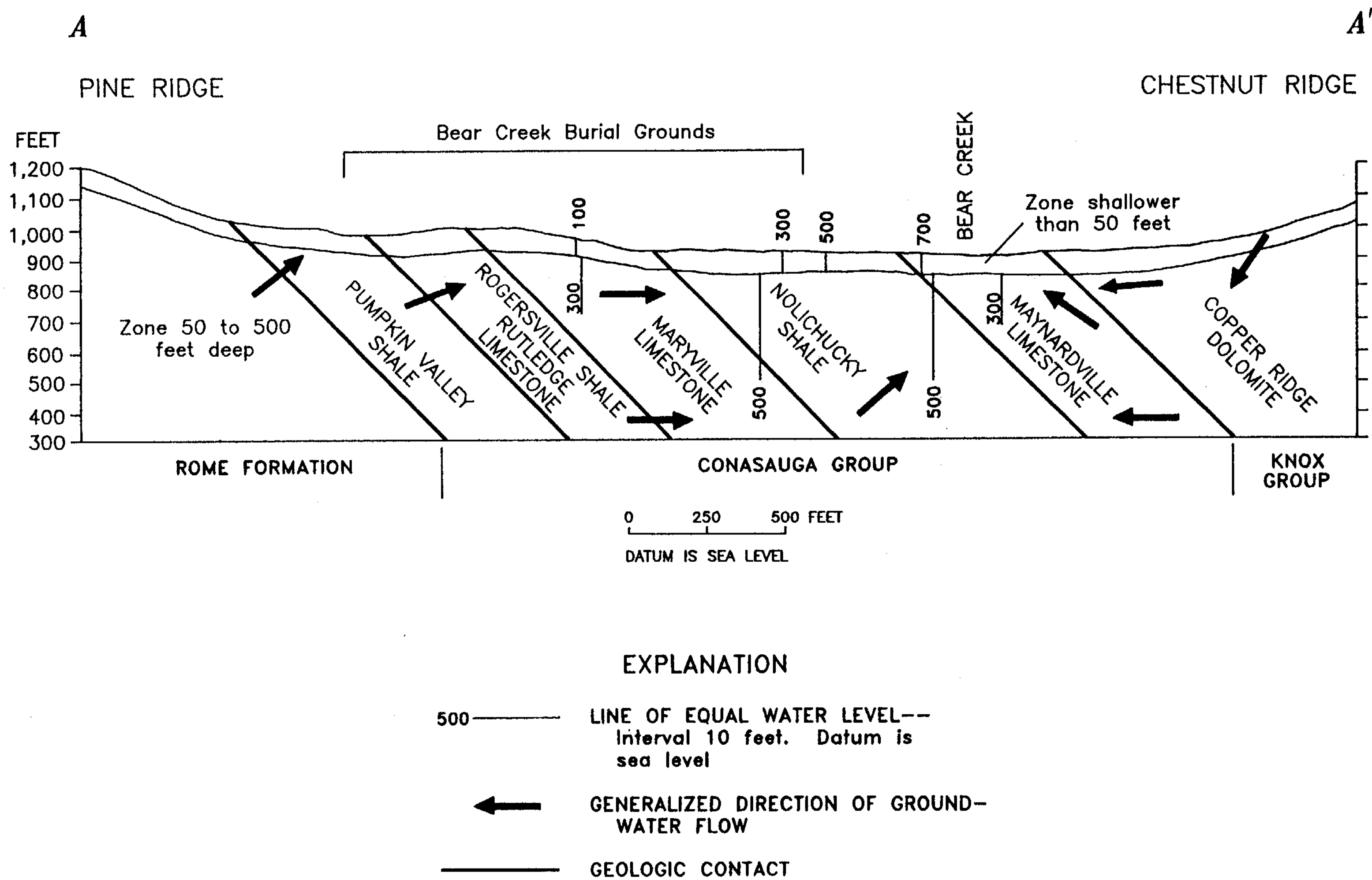


Figure 19.--Distribution of dissolved solids in the water-bearing zone shallower than 50 feet and the zone deeper than 50 feet (trace of section shown on figures 14 and 17).

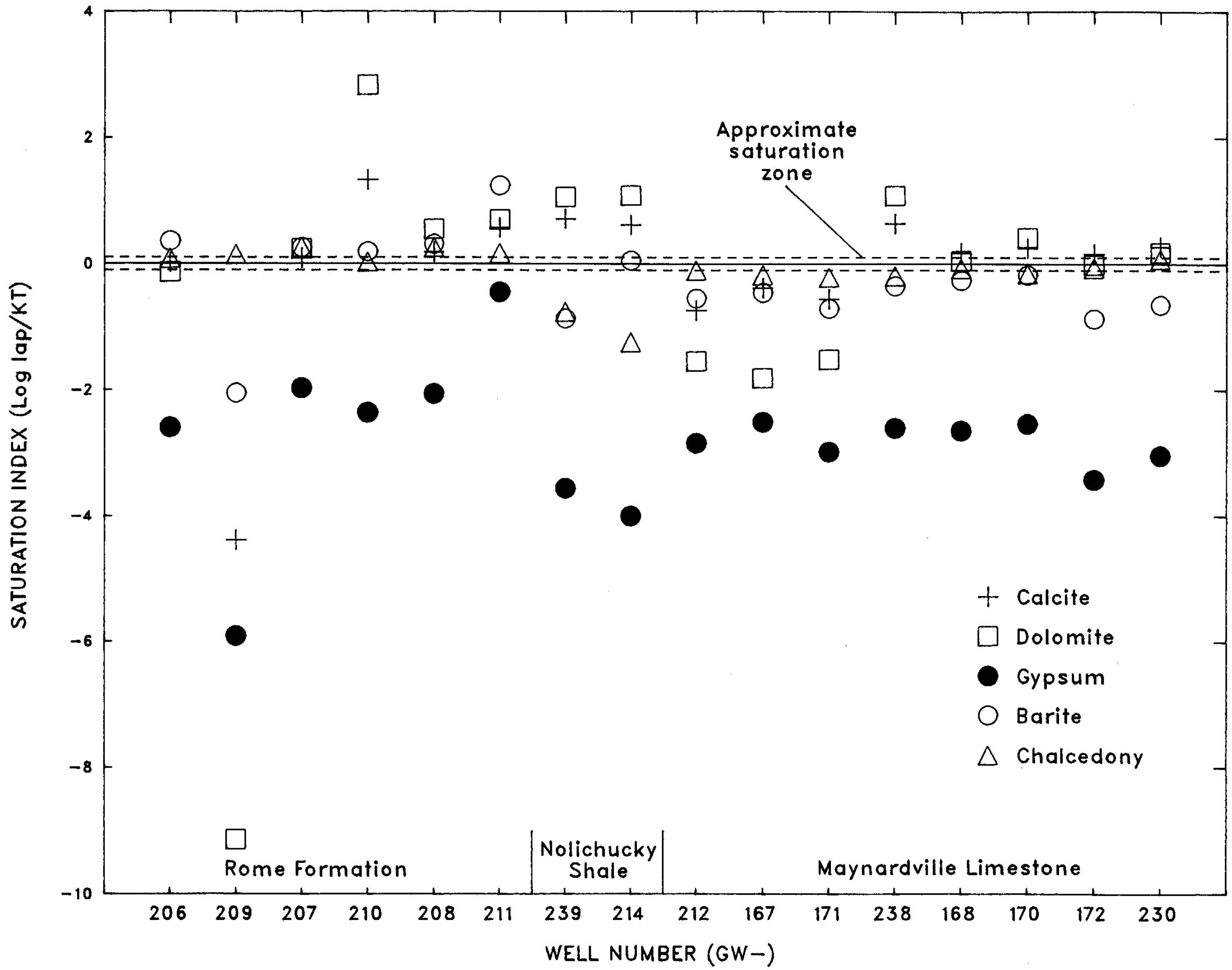


Figure 20.--Mineral saturation states of ground water from wells in the Rome Formation, Nolichucky Shale, and Maynardville Limestone. (Well locations shown on figure 3.)

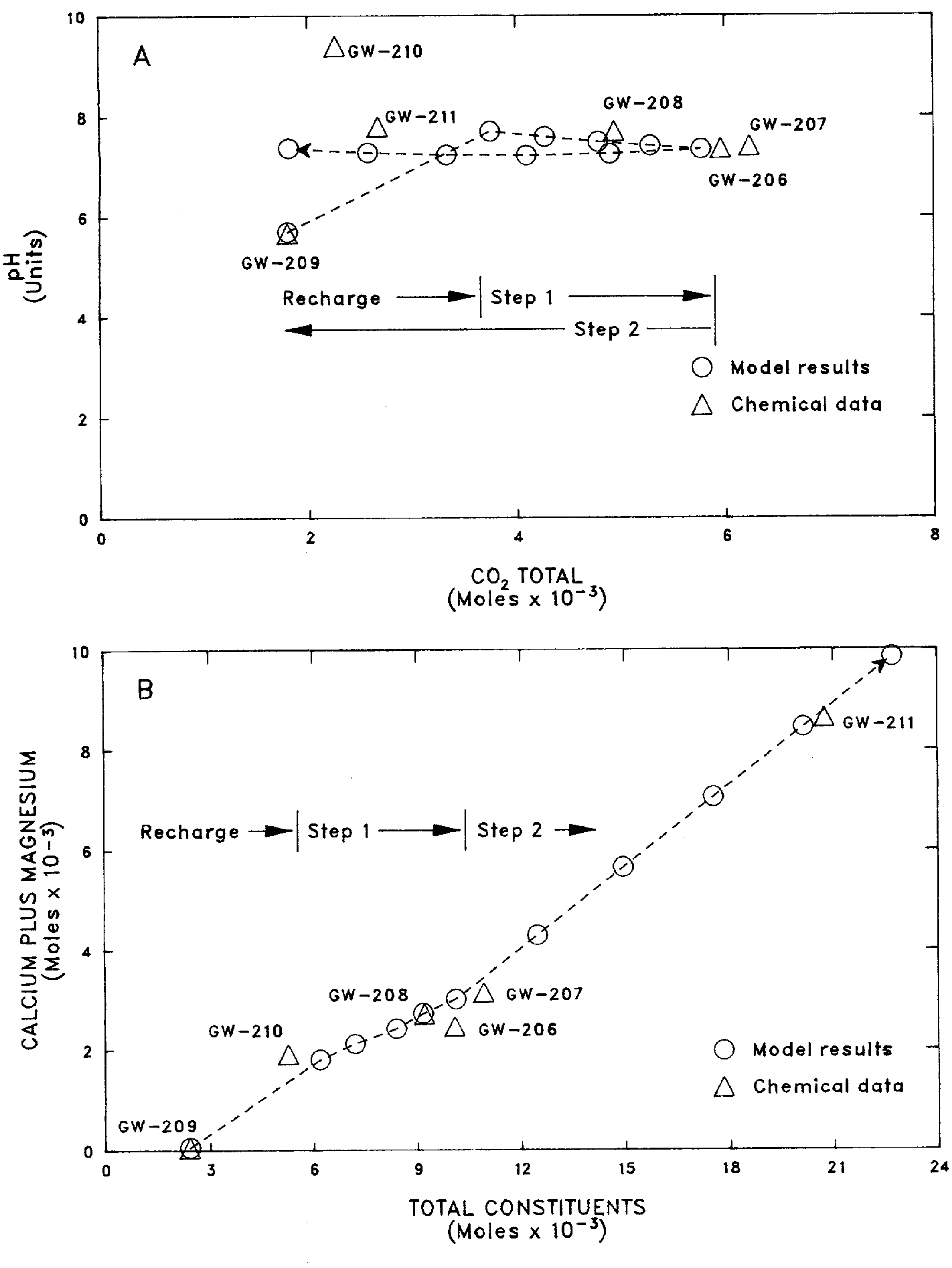


Figure 21.--(A) Chemical evolution of water in the Rome Formation from the PHREEQE model and (B) chemical analyses of water from selected wells. (Well locations shown on figure 3.)

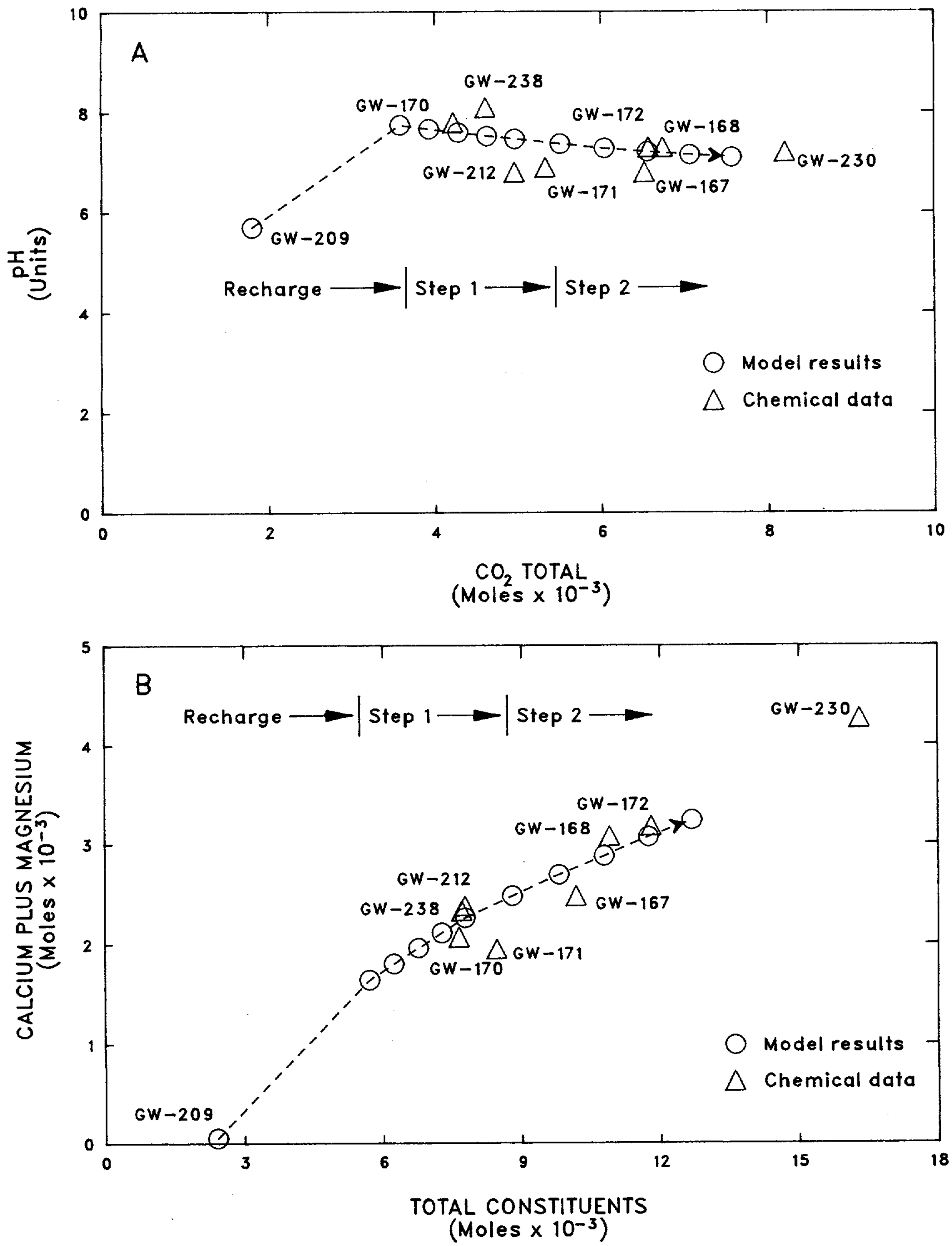


Figure 22.--(A) Chemical evolution of water in the Maynardville Limestone from the PHREEQE model and (B) chemical analyses of water from selected wells. (Well locations shown on figure 3.)

amounts dissolved in both models for the Rome Formation and the Maynardville Limestone. In order to account for excess chloride concentrations in the two analyses and for residual cations in the mass transfer, a mixed chloride salt containing sodium, potassium, and magnesium cations was used. The source of this compound or mix of compounds could be from the road salts used during winter months for snow removal. Elevated concentrations of dissolved chloride in wells GW-172 and GW-230 indicate that water from the Copper Ridge Dolomite may mix with water in the deeper parts of the Maynardville Limestone in the vicinity of Bear Creek. These chemical data were sparse and incomplete for the Copper Ridge Dolomite, and the mass transfer model is simplistic. Thus, the model is inadequately constrained for testing these hypotheses.

SIMULATION OF GROUND-WATER FLOW

MODEL ASSUMPTIONS

The finite-difference model of McDonald and Harbaugh (1988) was used to simulate the three-dimensional flow system in the regolith and bedrock. The following simplifications and assumptions were made to simulate the complex hydrologic system:

1. Fracture and solution zones are extensive enough in both areal and depth distribution that the regolith and bedrock can be simulated as porous media.

2. The top layer, representing the regolith and upper zone of weathered bedrock, is unconfined and deeper layers are confined.

3. The bottom of the model is 600 feet below the water table. The bottom is assumed to be a no-flow boundary, because of hydraulic-conductivity values several orders of magnitude lower than values measured at shallower depths.

4. The hydraulic characteristics of the geologic units are homogeneous within a block of the finite-difference grid.

5. The grid is aligned with primary axes of hydraulic-conductivity tensors and any anisotropy in a layer is uniform within that layer.

6. Flow within a layer is horizontal; flow (leakage) between layers is vertical.

7. The ground-water system is at steady state.

CONCEPTUAL MODEL

The regolith and bedrock hydrologic system was divided into four layers to simulate ground-water flow (fig. 23). The layers were determined on the basis of differences in physical characteristics that affect transmissivity, on the consistency of potentiometric data within a layer, and on the difference in potentiometric data (vertical gradient) between layers. Layer 1, 50 feet in thickness, corresponds to the saturated regolith and upper zone of weathered bedrock to which the regolith is hydraulically connected. This layer also corresponds to the upper, chemically distinct zone identified in the geochemical analysis. Layer 2, 50 feet in thickness, is the upper bedrock zone that is weathered and fractured. Layers 1 and 2 are hydraulically well connected with vertical flow between layers, and have virtually the same hydraulic properties (table 1) because of fractures that are parallel and normal to bedding. Layer 3, 300 feet in thickness, is characterized by fewer and smaller fractures and cavities, and thus the hydraulic conductivity is lower (table 1). Layer 4, 200 feet in thickness, has significantly lower hydraulic conductivity than the shallower layers.

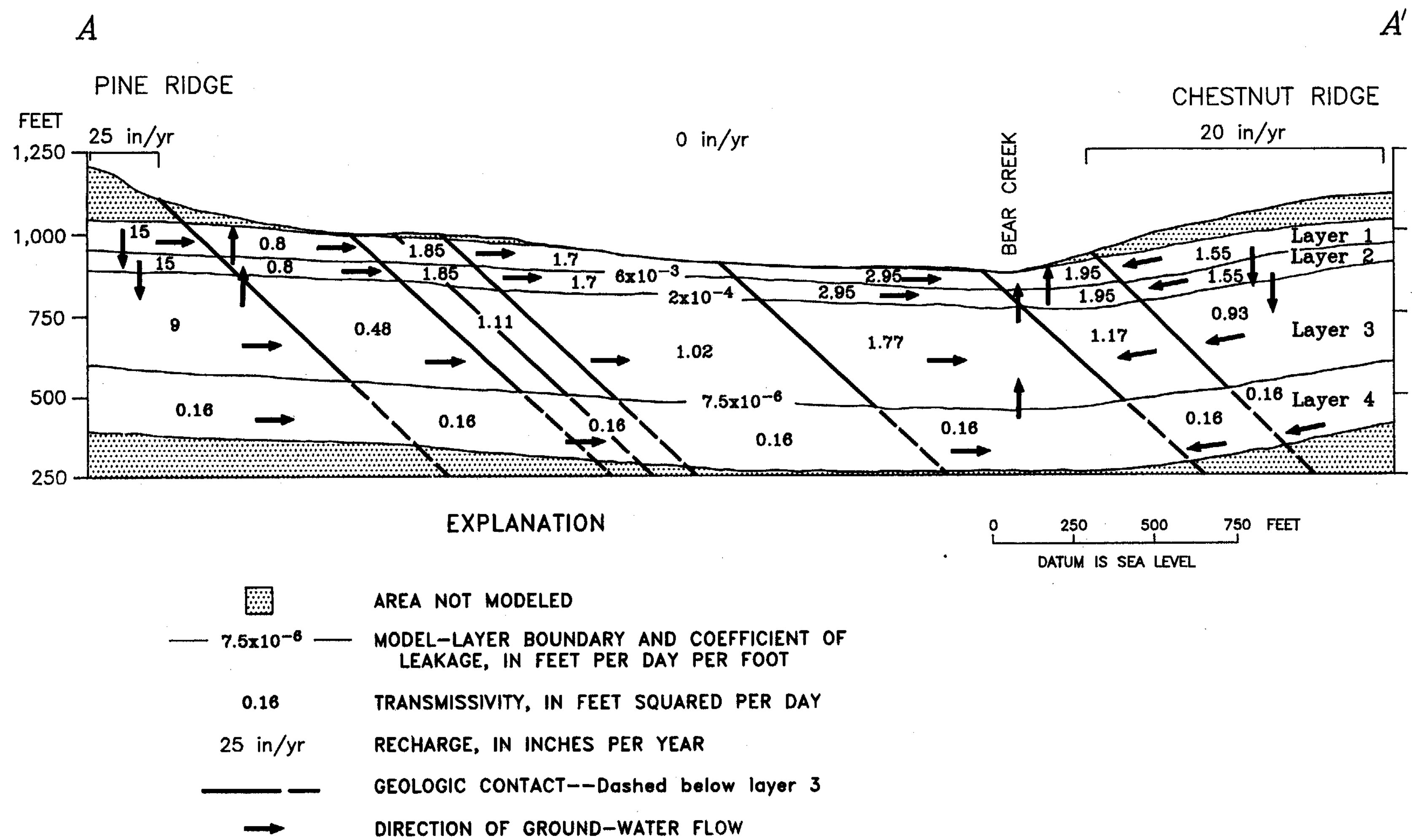


Figure 23.--Design of the digital flow model (trace of section shown on figure 1).

Layer 1 is unconfined. Layers 2, 3, and 4 are simulated as confined; however, there are no confining layers that cut across the geologic bedding in the real system. The dipping beds of the geologic units were not directly simulated, that is, the plane of the model is horizontal. However, offsets in the units with depth (fig. 23) were accounted for by offsetting the hydraulic conductivity of a layer toward Chestnut Ridge.

Bear Creek, East Fork Poplar Creek, Grassy Creek, Scarboro Creek, and the unnamed creek in Union Valley are assumed to be hydraulically connected to layer 1 through leaky streambeds. Elevations of the water surface of the simulated streams were assumed to be constant within a grid block, and ground-water gain or loss through the streambed was simulated. The Clinch River was assumed to maintain a constant head in layer 1 that is the same as the river stage, because it is virtually a controlled lake at each end of the valley. The ephemeral streams that cut across strike and springs are considered to be drains that can gain water from, but not lose water to, the ground-water system.

All recharge (distribution shown on fig. 23) is from precipitation and is primarily beneath the ridges. The system receives no subsurface recharge from outside the model boundaries. The valley floor receives no net recharge in most areas or discharges water locally (to streams and springs).

MODEL BOUNDARIES

The lateral boundaries of the model correspond to real hydrologic boundaries. The Clinch River forms both the eastern and western boundaries, and no underflow is assumed. Pine and Chestnut Ridges, the northern and southern boundaries, respectively, are drainage divides as well as ground-water divides. The upper boundary is the water table. The bottom boundary was set at a depth of about 600 feet below the water table and ranges between elevations of about 250 to 400 feet above sea level. The bottom boundary is assumed to be impermeable because of greatly decreased hydraulic conductivity and few secondary permeability features at those depths.

MODEL CONSTRUCTION

The 15-mi^2 grid of the model, encompassing the entire study area (fig. 1), is approximately a 1- by 15-mile rectangle consisting of variable size blocks (fig. 24). The smallest blocks are 250 by 500 feet and the largest are 250 by 1,000 feet.

Input for layer 1 included initial estimates of water levels, average transmissivity for each geologic unit (fig. 23), and recharge. Initial water levels were obtained from the potentiometric map that generally represents average water levels for October 1986 (fig. 8). This period represents seasonally low, steady-state ground-water levels, and probably represents a lower than average period. Antecedent precipitation for the previous year was 19 inches lower than average (fig. 7). A representative hydraulic-conductivity value, which was derived from statistical analyses and regression modeling (Connell and Bailey, 1989), was assigned to each geologic unit (table 1). Transmissivity was calculated from the representative hydraulic conductivities and a uniform thickness of 50 feet. Recharge rates applied to layer 1 were uniform within a geologic unit, and recharge was applied only to Pine and Chestnut Ridges (fig. 23). The initial recharge rates applied to Pine Ridge and to Chestnut Ridge were 25 and 20 in/yr, respectively, and were based on the results of cross-sectional model analyses described by Bailey (1988). Evapotranspiration was not simulated because its effect was included in the recharge rates.

The main streams were simulated as river nodes in layer 1 (fig. 24). Conductance (C), used to simulate leakage to and from river nodes, was calculated by:

$$C = \frac{KA}{b}$$

where
- K is vertical hydraulic conductivity of the streambed, in feet per day;
- A is the area of the river within the node, in square feet; and
- b is the streambed thickness, in feet.

Thickness of the streambeds was assumed to be 1 foot to simplify calculations; a vertical hydraulic conductivity of 1 ft/d was used initially for all streams. These initial values could be changed during calibration if simulated seepage to the streams did not approximate measured seepage. The streambed bottom within each river node is the elevation of the stream on a topographic map, and stream stage was calculated for each node assuming a 1-foot water depth. Stage of dry stream reaches was the same as streambed elevation to minimize leakage.

The Clinch River was simulated as a constant-head boundary in layer 1 at each end of the model (fig. 24). Stage of the Clinch River was set at the average stage during October: 794 feet above sea level on the east end, and 740 feet on the west end.

Tributaries that flow across strike were simulated as drains to the ground-water system (fig. 24). Elevation of the drains is the elevation of the stream channel obtained from topographic maps. Conductance of the drain bottoms was calculated by:

$$C = \frac{Q}{^\wedge h}$$

where
- Q is discharge, in cubic feet per second; and
- $^\wedge h$ is the difference in head between the water table and the drain, in feet.

Conductance between the tributaries and the rock units was unknown, but discharge measured during base-flow conditions was assumed to equal ground-water gain in the streams. The maximum difference in head between the water table and the drain was assumed to be 1 foot, and so, conductance equalled the discharge value. Discharge at the mouth of the tributary measured in August 1985 was used where available. April 1984 data were used for tributaries that were not measured in August 1985. In order to make conductance values calculated from April streamflow data compatible with the August data, the conductance values were generally reduced by an order of magnitude to account for the order-of-magnitude difference in discharge at the gage between the April and August measurements. The calculated conductance value for each tributary was applied to each drain node representing that tributary. Average values from nearby measured tributaries were applied to those tributaries that were not measured. Because natural, pre-construction drainage still affects ground-water flow beneath the Y-12 Plant, drain nodes were put in that area based on Rothschild and others (1984, p. 26, 35).

Springs along the contact between the Maynardville Limestone and Copper Ridge Dolomite in the Bear Creek watershed were simulated as drains in layer 1 (fig. 24). Their conductance values were also assumed to be equal to the measured discharge value.

Input for layers 2, 3, and 4 consisted of estimates of initial water levels and transmissivity for each layer. Initial water levels for layer 2 were from a potentiometric map constructed from water levels that generally represent October 1986, and transmissivity (fig. 23) was initially set at the same value as transmissivity in layer 1. Initial water levels for layers 3 and 4 were the same as for layer 2, because water-level measurements were too sparse to construct a complete potentiometric map for layers 3 and 4. Transmissivities for layer 3 (fig. 23) were calculated

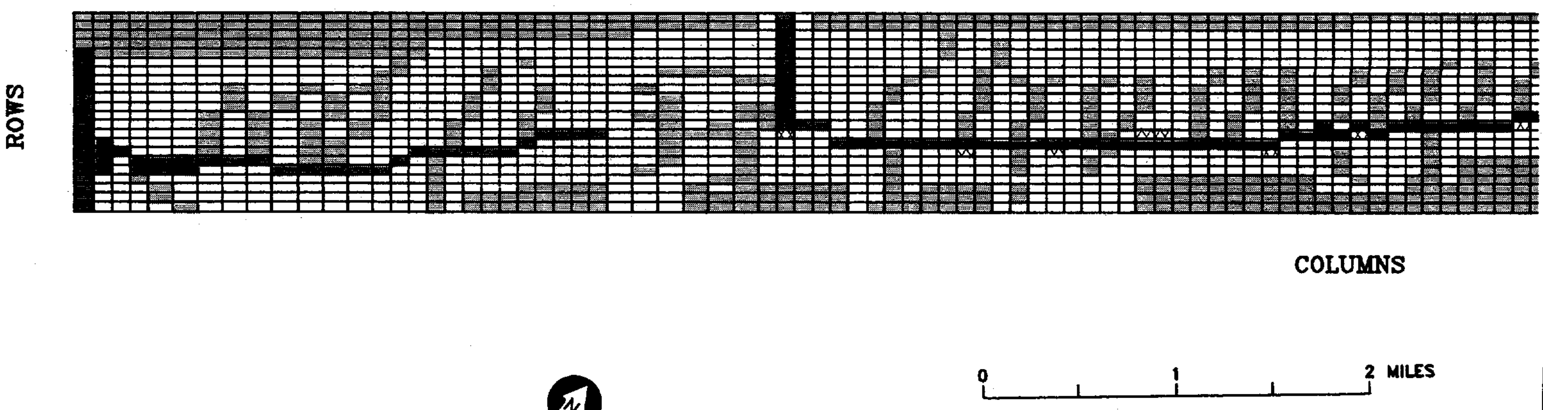


Figure 24.--Finite-difference grid for the digital flow model.

using a thickness of 300 feet and hydraulic-conductivity values that were one order of magnitude lower than the values used for geologic units in layers 1 and 2 (table 1), because fractures and solution features are less prominent below the bottom of layer 2. Transmissivity in layer 3 was offset by the width of one node (250 feet) toward Chestnut Ridge to approximate the downdip shift of formations at depth. Hydraulic conductivity for layer 4 was the same value (7.8×10^{-5} ft/d) for all geologic units, and a transmissivity value was calculated using a thickness of 200 feet (fig. 23).

Leakage between model layers was simulated by vertical conductance. Because the layers were assumed to be hydraulically well connected and not separated by confining material, vertical conductance between layers was calculated using the aquifer properties. Vertical conductance is calculated within the model using values of vertical leakance (McDonald & Harbaugh, 1988, p. 5-11). Vertical leakance (VL) between adjacent layers was calculated by:

$$VL = \frac{2K_{La} \; K_{Lc}}{K_{La} \; b_{Lc} + K_{Lc} \; b_{La}}$$

where
- K is vertical hydraulic conductivity, in feet per day;
- b is thickness, in feet;
- La is the uppermost layer; and
- Lc is the lowermost layer.

In order to calculate the largest reasonable vertical leakance between model layers for initial runs, the highest hydraulic-conductivity value (the value for the Rome Formation in each layer) was used to calculate a leakance value that was applied uniformly between the model layers (fig. 23). Calculations were based on the initial estimates of hydraulic conductivity (table 1). Vertical leakance between layers 1 and 2 was 6.0×10^{-3} (ft/d)/ft, between layers 2 and 3; 2.0×10^{-4}, and between layers 3 and 4, 7.5×10^{-6} (fig. 23).

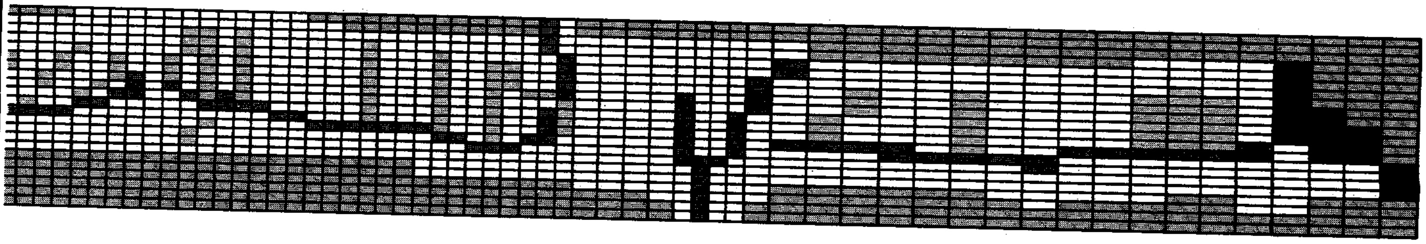


Figure 24.--Finite-difference grid for the digital flow model--Continued.

EXPLANATION

CONSTANT HEAD

RIVER

DRAIN

SPRING (SIMULATED AS DRAINS)

INACTIVE NODE

MODEL CALIBRATION

The ground-water flow model was calibrated to water levels averaged for October 1986, and to the range of average ground-water seepage per mile of stream channel. The system is assumed to have been at steady state at that time. Of the 132 water levels used for calibration, 69 percent were from wells measured weekly during October (63 were measured during October 1986, 28 were measured during October of other years). The remaining 41 wells (31 percent of the total) were not measured during or had not yet been drilled by October 1986, so the lowest available water levels were used. Seventy-three percent of the water levels were measured during 1986. Water levels were used for comparison in 62 nodes for layer 1, 9 for layer 2, 20 for layer 3, and 11 for layer 4. The range of average ground-water seepage was calculated from streamflow during low and high base flow conditions. Seepage for the October calibration period was probably at the low end of the range due to low recharge over the summer months and low ground-water levels.

Water levels from an additional six wells in the Grassy Creek watershed were available for calibration for layers 1, 2, and 3; however, their water levels were not compatible for the calibration period represented by water levels in the other wells. The most recent measurements from these six wells were from October 1983 or 1984. Antecedent precipitation for those years was 5 and 20 inches greater, respectively, than precipitation in 1986. Therefore, these water levels were omitted from calibration and the simulated water levels in the area were accepted.

Transmissivity for all formations in each layer, leakage coefficients between layers, recharge, and horizontal anisotropy of hydraulic conductivity were varied during the calibration process to maximize matches between simulated- and measured-head values and simulated and measured ground-water seepage. Following calibration, no substantial differences between calibrated and initial values were noted with the exception of recharge rates and local values of transmissivity for the Maynardville Limestone in layers 1 and 2. Following are discussions of selected comparisons of model results that document the calibration process and that may provide insight into model uncertainty and parameter sensitivity.

Comparison between model results without the tributaries simulated as drains and results with the tributaries simulated as drains demonstrated the importance of the normal-to-strike tributaries to the flow system in the valley. Simulations without the drains produced heads that were hundreds of feet higher than measured heads, especially on the ridges, but simulations with the drains produced heads that were closer to measured heads and the conceptualized flow system. However, a reasonable match to measured heads required a reduction of the rates of recharge applied to Pine Ridge and Chestnut Ridge to 5.0 and 4.0 in/yr, respectively. These values, which are about 20 percent of initial estimates, are in general agreement with recharge rates used in the simulation of ground-water flow in a comparable hydrogeologic setting (Tucci, 1986).

Simulation of horizontal anisotropy in hydraulic conductivity that is greater parallel to strike than normal to strike had little effect on improving head matches during calibration and did not substitute for the control these tributaries have on short flowpaths along strike.

Measurements in piezometers installed upgradient from the Burial Grounds near the Rome Formation and Pumpkin Valley Shale contact show a steep upward gradient. In simulations using the initial transmissivity values for the Pumpkin Valley Shale (fig. 23), head matches were good for the well in layer 1 but were poor for the well in layer 3. Several combinations of transmissivity of the Pumpkin Valley in each layer were tested to improve head matches for these wells. Head matches for both wells were made much worse by making the hydraulic conductivity of the Pumpkin Valley Shale in layer 3 the same as in layers 1 and 2. However, because of the greater thickness of layer 3, the effect of making conductivity the same was to raise transmissivity of the Pumpkin Valley Shale in layer 3 (from 0.48 to 4.8 ft^2/d) higher than in the upper layers (0.8 ft^2/d). Other combinations of conductivity in the Pumpkin Valley, which (1) lowered the conductivity in layer 3 up to two orders of magnitude or (2) lowered the conductivity in layer 2 by one order of magnitude, improved head matches for the well in layer 3 but caused worse head matches for the well in layer 1. No combination of transmissivity successfully matched heads in both layer 1 and layer 3. For all cases, model-simulated heads for the well in layer 3 were 30 to 50 feet lower than measured head, which may indicate that an additional source of water is needed in the model that is not included in the conceptual model. However, no head or flux data are available for estimating water entering the system at depth under Pine Ridge. The poor head match for the well in layer 3 for all combinations of conductivity in the Pumpkin Valley Shale also may be caused by very localized high-permeability conditions that are not simulated at the scale of this model. These changes in conductivity of the Pumpkin Valley Shale affected few other simulated heads. The initial values of conductivity for the Pumpkin Valley Shale were retained because that combination produced the best head matches for those two wells.

Localized changes in transmissivity of a formation were made in only one area. Transmissivity of the Maynardville Limestone in the Scarboro Creek watershed was increased by one order of magnitude in layers 1 and 2. This adjustment improved head matches considerably, and was considered justifiable in this location because a wide lineament that extends for several miles passes almost normal to strike through that part of the valley. The lineament is probably an expression of a major fracture or fault zone that caused increased secondary permeability in the formations.

Improved agreement between simulated and measured heads in layer 4 were achieved during calibration by any changes in layer or leakage characteristics that allowed more water to enter layer 4 from layer 3. Although these changes improved the head matches in layer 4, matches in layer 3 were made worse. This result may indicate that layer 4 needs a source of water from the ridges, as well as from overlying strata. However, no data on heads or influx of water through the lateral boundaries were available.

The rates of recharge at Pine Ridge and Chestnut Ridge derived from calibration of the areal model described in this report (4.0-5.0 in/yr) are substantially lower than corresponding rates determined during the cross-sectional model analysis (20-25 in/yr; Bailey, 1988). Most of this disparity can be attributed to differences in nodal resolution between the two models. The same lateral distance from ridge line to ridge line across Bear Creek Valley is discretized using 68 nodes for the cross-sectional model and 22 nodes for the areal model. Accordingly, the cross-sectional model accounts for a greater percentage of total ground-water flow than the areal model. The additional flow probably represents discharge from the local flow regime (Toth, 1962) that, for this area, may be analogous to flow in the "stormflow zone" described by Moore (1988). Differences between mean (normalized) recharge rates used in the areal model (1 in/yr) and in the cross-sectional model (14 in/yr) can be similarly accounted for.

Model-simulated water levels for each layer (fig. 25) are considered to be a good representation of the overall flow system, even though some of the steep gradients under the ridges could not be matched. Although simulated water levels are similar for all the layers, the patterns of vertical flow between the layers differ greatly (fig. 26). Flow between layers 1 and 2 is downward beneath the ridges, and upward at the break between Pine Ridge and the valley floor and beneath streams. Flow in layers 1 and 2 is essentially horizontal across the valley floor where there is no vertical flow between layers. Flow between layers 2 and 3 is downward beneath the ridges and flow beneath the valley floor is both upward and downward. The flow pattern between layers 3 and 4 also shows downward flow beneath the ridges and upward flow beneath the valley. These vertical-flow patterns are consistent with the conceptualization of the three-dimensional flow system. The combination of hydraulic characteristics in the model is not a unique solution for simulating the system, but the model is considered to be well calibrated to the available data.

Simulated ground-water seepage to the main streams ranges from 0.01 to 0.05 ft^3/s per mile of stream channel, which are within the range of average seepage per mile of stream channel (about 0 to 1 ft^3/s) calculated from stream measurements during low and high base flow conditions. The low values for simulated seepage reflect the calibration to low-flow conditions.

Components of the simulated water budget are summarized in table 6. Simulated no-flow conditions occur at the lateral boundaries, and only 5 percent of total recharge discharges directly to the Clinch River. Streams (other than the Clinch River) are the primary drains for the system; the streams that are subparallel to the valley axis and strike receive 72 percent of the ground-water discharge, and small tributaries that are normal to strike and springs (modeled as drains) receive 23 percent. Areal recharge provides 97 percent of total ground water and leakage from losing reaches of streams provides 3 percent.

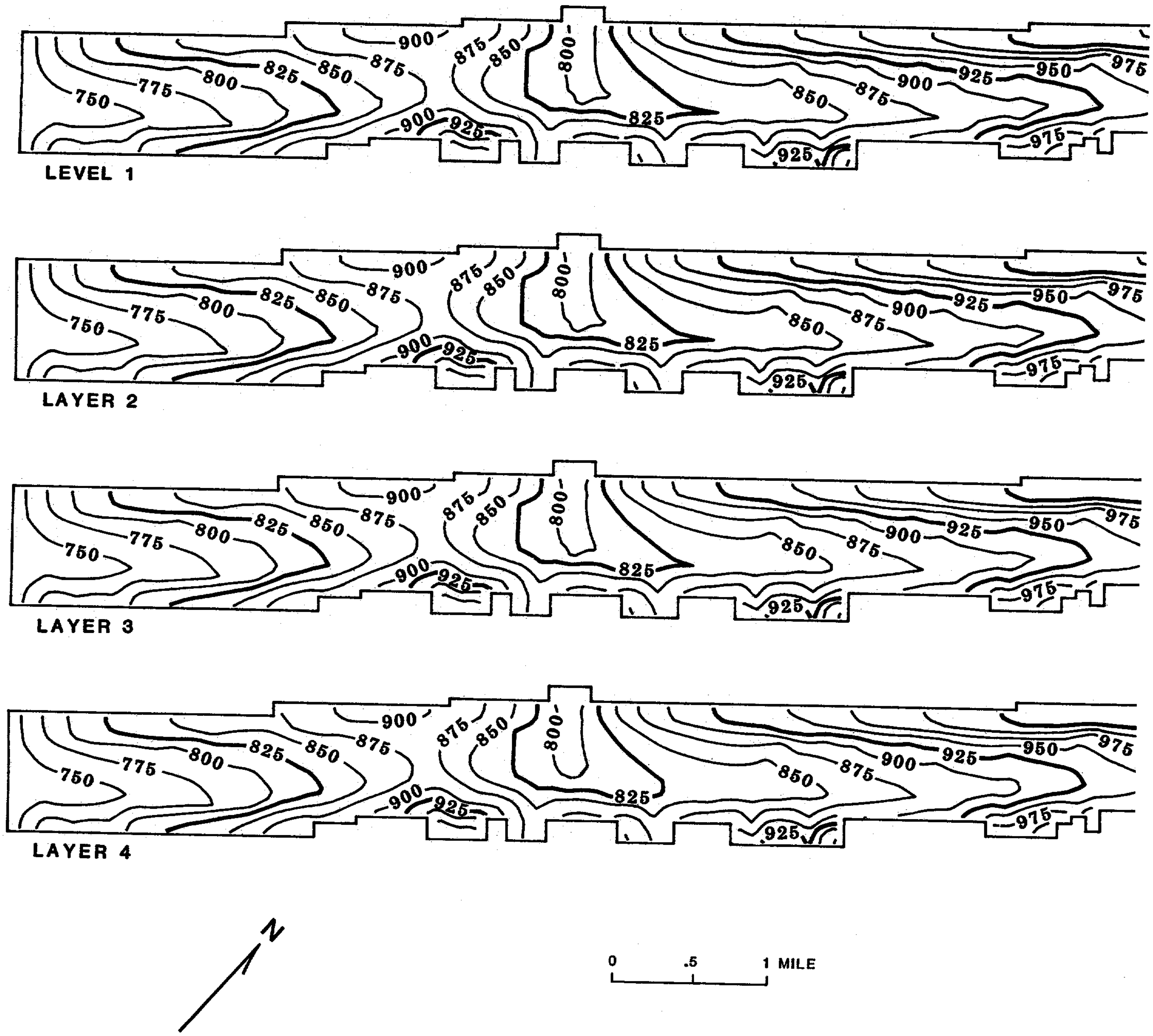
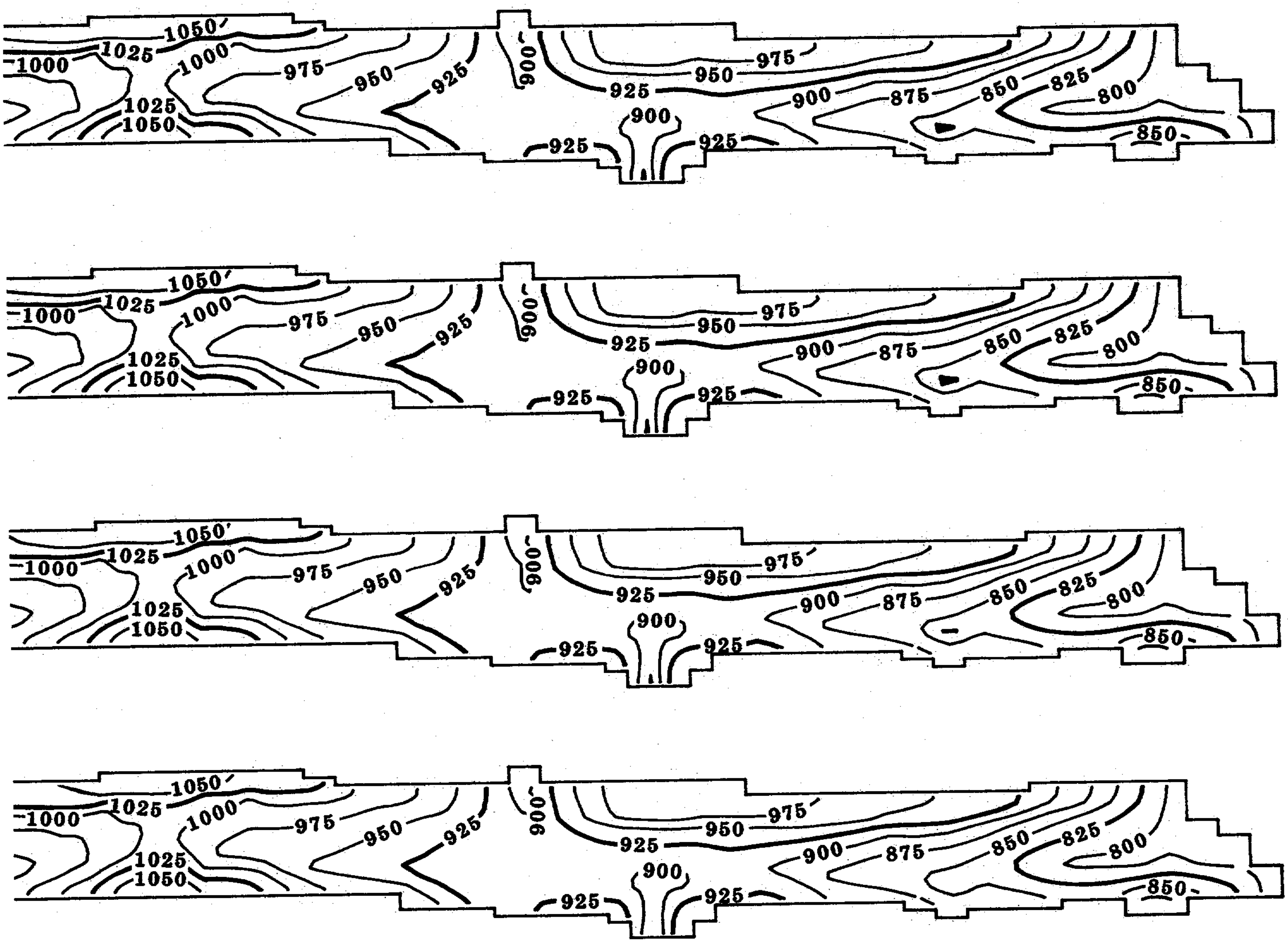


Figure 25.--Model-simulated water levels in layers 1, 2, 3, and 4.



EXPLANATION

—925— LINE OF EQUAL WATER LEVEL—INTERVAL 25 FEET. DATUM IS SEA LEVEL

——— ACTIVE-NODE BOUNDARY

Figure 25.--Model-simulated water levels in layers 1, 2, 3, and 4--Continued.

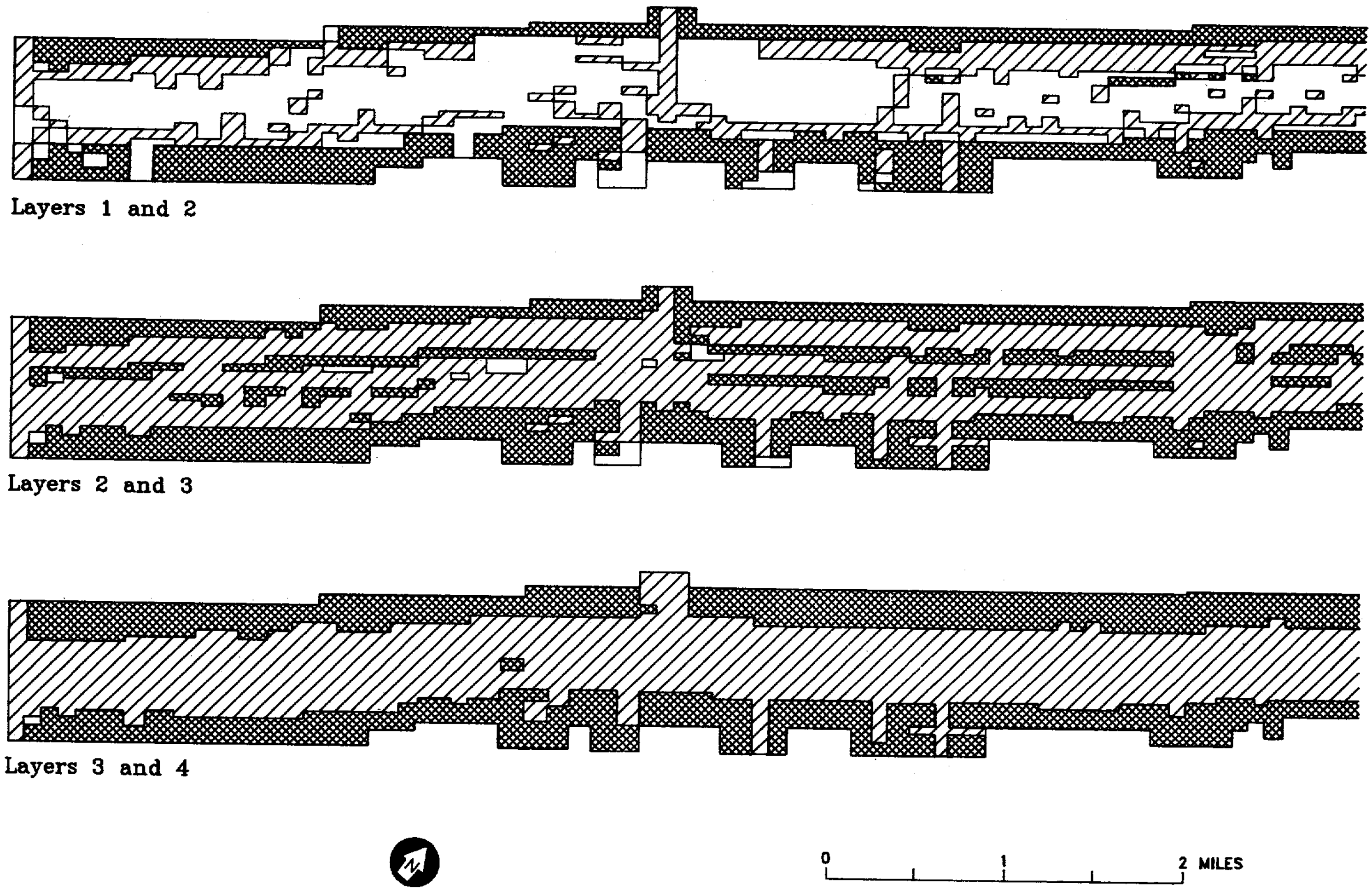


Figure 26.--Leakage direction between layers 1 and 2, 2 and 3, and 3 and 4.

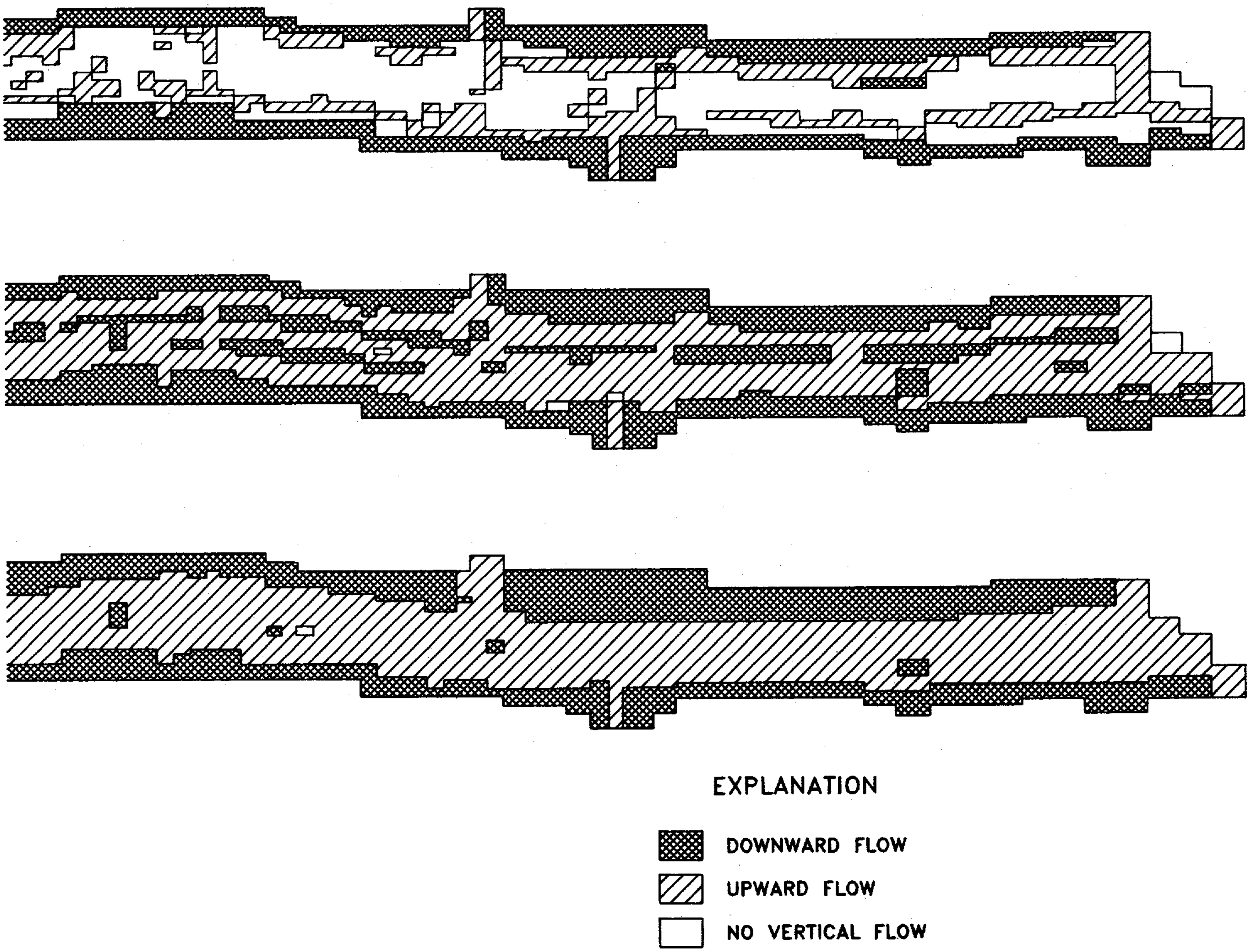


Figure 26.--Leakage direction between layers 1 and 2, 2 and 3, and 3 and 4--Continued.

Net leakage between layers is about zero; however, about 0.4 ft^3/s is exchanged (both downward and upward) between layers 1 and 2, about 0.2 between layers 2 and 3, and about 0.002 between layers 3 and 4. The relative rate of water exchanged between adjacent layers is an indication of the greater activity of the shallower flow system compared to deeper flow. Recharge into layer 2 is 65 percent of the recharge to layer 1 (0.61 ft^3/s, table 6), recharge to layer 3 is 32 percent (and 0.5 percent of the recharge to layer 2), and to layer 4, 0.3 percent (and 0.01 percent of the recharge to layer 3).

SENSITIVITY ANALYSIS

The response of the model to adjustments in recharge, hydraulic conductivity of the layers, vertical conductance between layers, hydraulic conductivity of the streambeds, and horizontal anisotropy was evaluated using sensitivity analyses. Hydraulic conductivity of each formation in all layers was adjusted by the same multiple for each sensitivity test (rather than each layer being adjusted individually while the other three layers were held constant). All three leakage layers were also adjusted by the same multiple for each test, and horizontal anisotropy was adjusted by the same multiple for all model layers for each test. Hydraulic conductivity of all the streambeds (both river and drain nodes) were varied by the same multiple, and conductivity of rivers and drain nodes was varied separately to distinguish any sensitivity to flow along strike. Ranges over which the hydraulic characteristics were varied are summarized on table 7.

Table 6.—*Model-calculated, steady-state water budget for Bear Creek Valley for seasonally low conditions, October 1986*

Sources and discharges	Flow, in cubic feet per second	Percentage of total
Sources		
Clinch River	0	0
Areal recharge	.59	97
Leakage from streams	.02	3
Total	.61	100
Discharges		
Ground-water seepage to:		
Clinch River	0.03	5
Streams	.44	72
Drains	.14	23
Total	.61	100

Differences between measured and simulated water levels were used as indicators of the sensitivity of the model to adjustments of a variable. The root mean square error (RMSE) was calculated for measured and simulated water levels by:

$$\mathrm{RMSE} = \sqrt{\frac{\sum_{i=1}^{N} (h_i^m - h_i^c)^2}{N}}$$

where

N is the number of observations (132);
h_i^m is the measured water level, in feet; and
h_i^c is the calculated water level, in feet.

RMSE was plotted for each adjustment in a variable to display the range of sensitivity.

The overall RMSE for all layers in the calibrated steady-state model is 14.0 feet. The RMSE of each layer is 15.3 feet for layer 1, 10.2 feet for layer 2, 16.3 feet for layer 3, and 13.5 feet for layer 4. Average head difference between simulated and measured heads for the model is -1.7 feet and the standard deviation is 13.9 feet. Seventy-three percent of the simulated heads differ from measured heads by less than 10 feet, which is within the normal range of seasonal water-level fluctuation of most wells.

The model is very sensitive to adjustments in recharge, particularly to increases in recharge (fig. 27). Calibrated recharge is only 20 percent of the initial recharge, which was derived from the cross-sectional modeling, and the calibrated values are lower than the average recharge calculated by Evaldi and Hoos. The lower recharge rates in the calibrated model are more consistent with estimates by Tucci (1986) and Moore (1988) (see "*Recharge*"). This sensitivity may indicate that streambed conductance or hydraulic conductivity, particularly of layer 1, is not compatible with the actual (higher) recharge, that the recharge estimates are incorrect, or that another mechanism for dissipating recharge is not accounted for in the model. It is likely that the latter is the case and the "stormflow zone" proposed by Moore (1988), which results in lower calculated recharge, may be an explanation for the sensitivity of the model to high recharge rates. Different combinations of hydraulic conductivity would also affect the recharge rate; however, unreasonably high hydraulic-conductivity values would be necessary to support higher recharge rates. Both distribution and rate of actual recharge to the ground-water system are difficult to evaluate due to indirect estimation techniques. Because model-calculated seepage to streams is in the range of measured seepage, the combination of recharge and hydraulic conductivities for the calibrated model is considered reasonable.

The model is very sensitive to adjustments in hydraulic conductivity of the layers (fig. 28). Changes in individual layers might not have the same effect on the model, but a unique combination of conductivities among the layers would be difficult to determine. Sensitivity is greater to decreases in conductivity because of the difficulty in transmitting water (recharge remains at

Table 7.—*Ranges of variation of hydraulic characteristics for sensitivity analyses*

[in/yr, inch per year; ft^2/d, foot squared per day; (ft/d)/ft, foot per day per foot]

Hydraulic characteristic	Calibrated value		Range of variation
Recharge (in/yr)	5.0 on Pine Ridge 4.0 on Chestnut Ridge		0.5 to 25.0 .2 to 20.0
Transmissivity of layers (ft^2/d)	see figure 20		0.1 to 100 times the value of each formation
Coefficient of leakage between layers: [(ft/d)/ft]	1 and 2 2 and 3 3 and 4	6×10^{-3} 2×10^{-4} 8×10^{-6}	6×10^{-5} to 6×10^{-1} 2×10^{-6} to 2×10^{-2} 8×10^{-8} to 8×10^{-4}
Conductance of streambeds and drains [(ft/d)/ft of channel]	0.09 to 18 1 to 35		9×10^{-5} to 1,800 1×10^{-4} to 350
Horizontal anisotropy[1]	1:1		10:1 to 1:1 1:1 to 1:5

[1]Ratio of 1:1, hydraulic conductivity is equal parallel and normal to strike. Ratios >1:1, conductivity is greater parallel to strike. Ratios <1:1, conductivity is greater normal to strike.

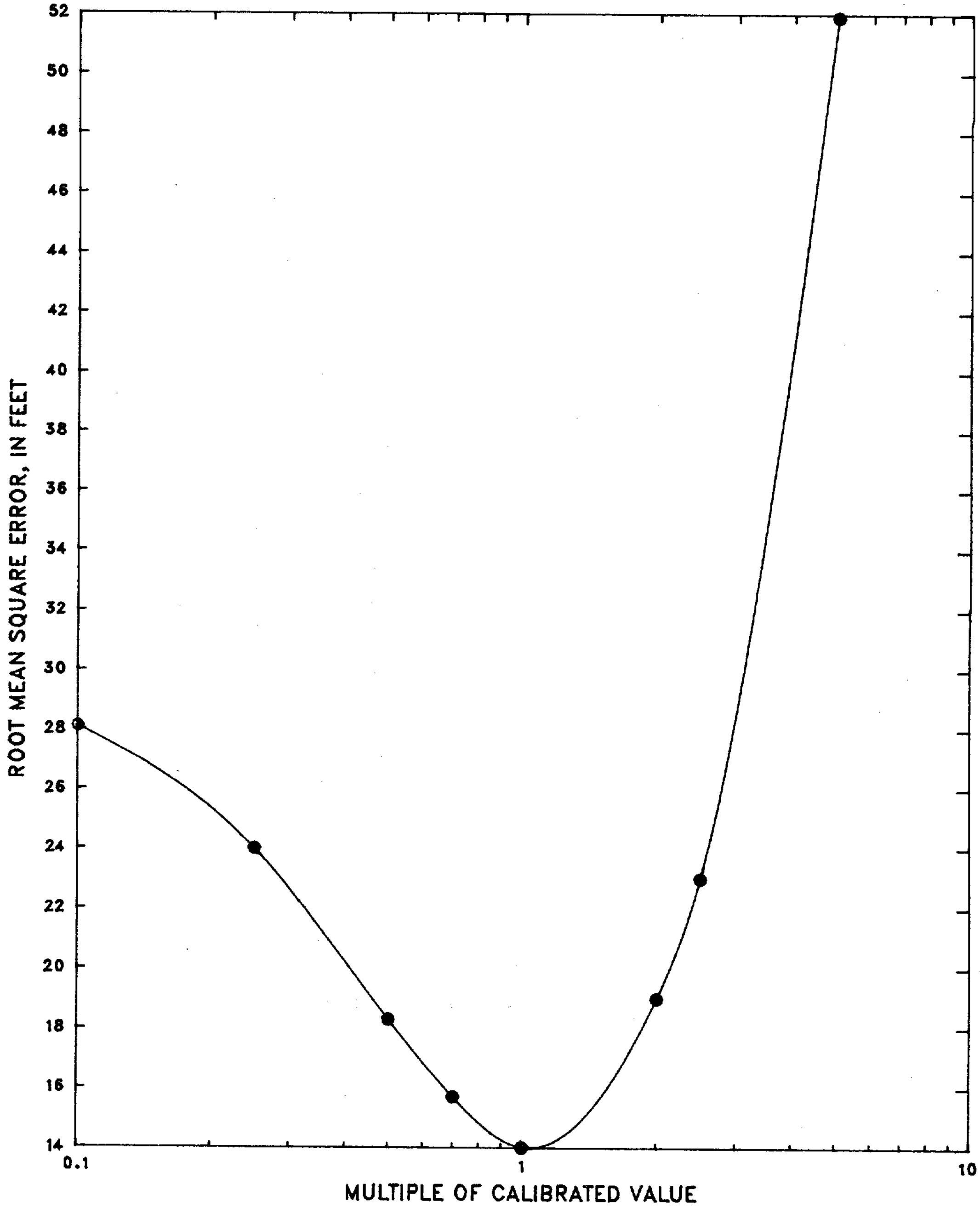


Figure 27.--Sensitivity of the digital flow model to adjustments in recharge.

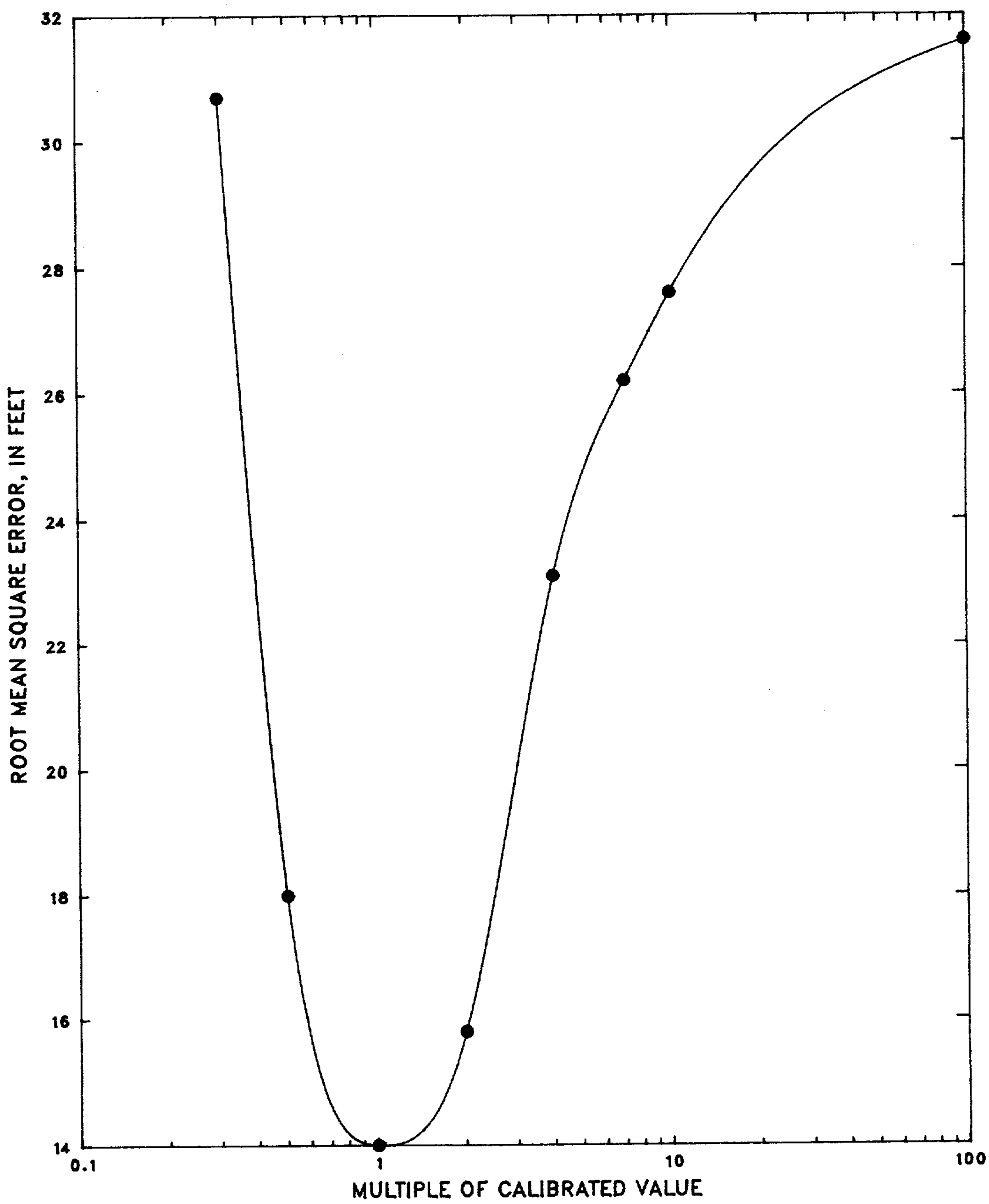


Figure 28.--Sensitivity of the digital flow model to adjustments in hydraulic conductivity of layers.

the calibrated value) through the system. If recharge and conductivity were varied in combination, a different, though not unique, combination could be determined that would match the head configuration. A limiting factor would be the rate of ground-water seepage to the streams.

The model is insensitive to a wide range of vertical conductance between the layers (fig. 29). Individual model layers would be more sensitive to adjustments in individual vertical-conductance layers.

The model is insensitive to increases in all streambed conductance values; however, the model becomes mathematically unstable for multiples of the calibrated values greater than 25 times (fig. 30). Neither is the model sensitive to as much as two orders of magnitude lower streambed conductivity values. A similar pattern of insensitivity results from variations in only river-node conductance values (fig. 31), but the model is more stable for the extreme multiples, which indicates the importance of ground-water seepage to the tributaries (drains). The model is completely insensitive to several orders of magnitude of variation in drain conductance only (fig. 31). However, the balance of the water budget is poor for multiples less than 0.001 and

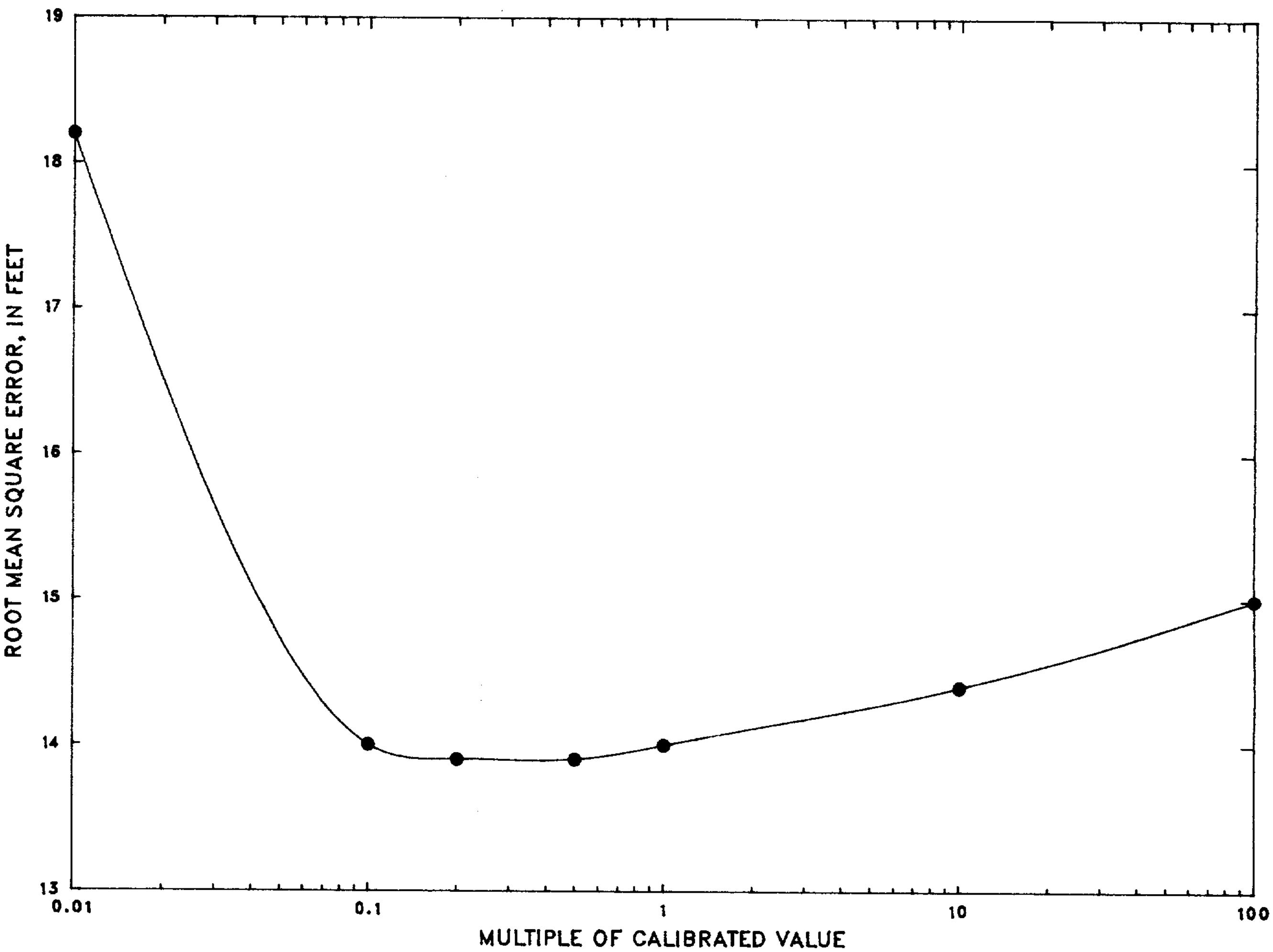


Figure 29.--Sensitivity of the digital flow model to adjustments in vertical conductance between layers.

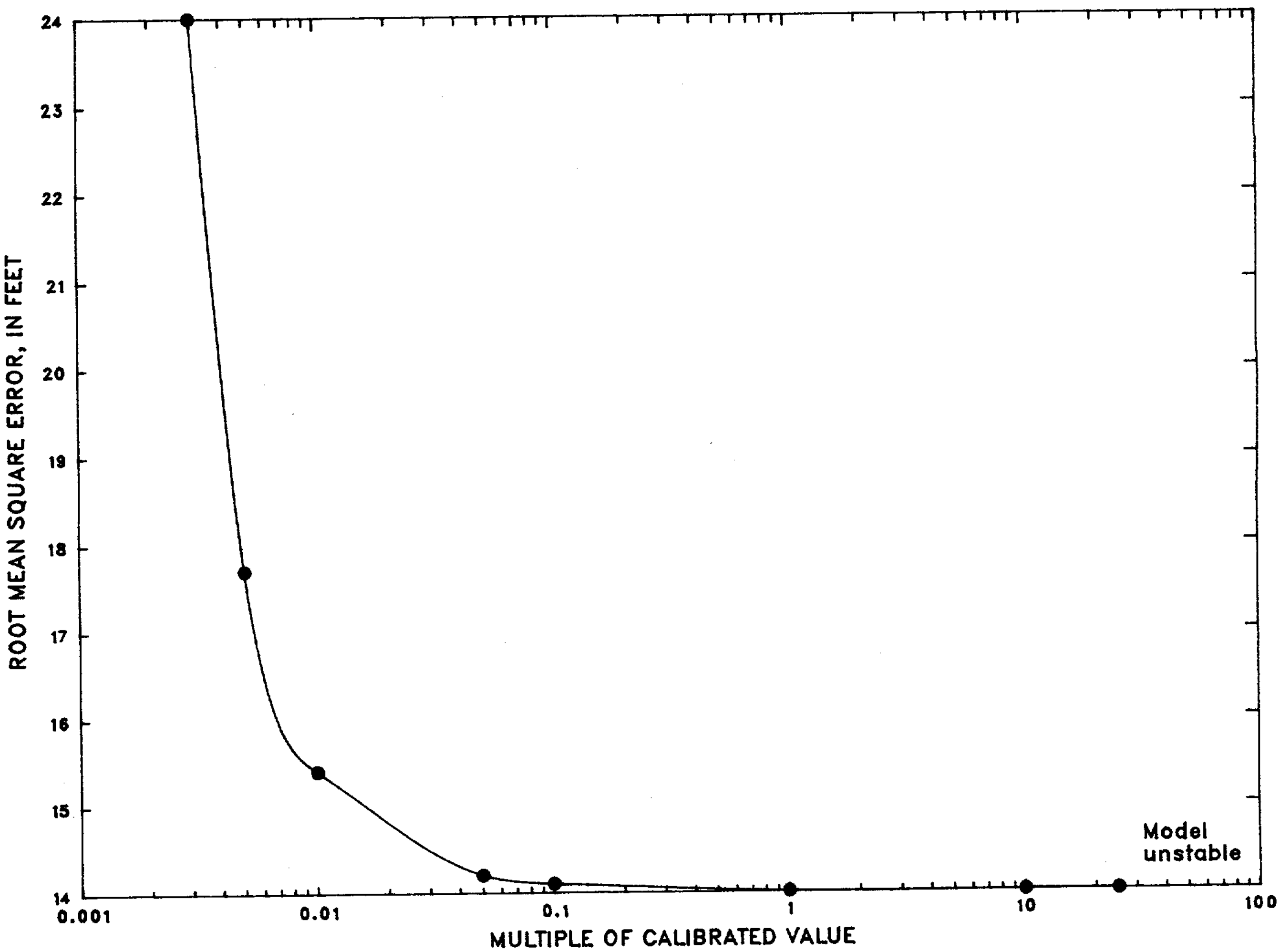


Figure 30.—Sensitivity of the digital flow model to adjustments in hydraulic conductivity of streambeds.

the model becomes unstable for multiples greater than 10 times the calibrated values. There is some improvement in the RMSE for drain conductances between 0.001 and 0.1 times the calibrated values. These improvements are localized and probably reflect local differences in streambed conductance of tributary streams; however, the simulation of the overall flow system is nearly the same as that of the calibrated model. The tributaries (drains) have less effect on the flow system than the main streams (river nodes), but the tributaries are essential to maintaining the water balance of the system.

Decreasing multiples of horizontal anisotropy correspond to ratios that favor flow along strike, and increases correspond to ratios that favor flow normal to strike (fig. 32). Ratios from 1.1 and 1.25 to 1 (favoring flow along strike) produced slightly better than calibrated RMSE values. The majority of head matches were about the same as in the calibrated model. Head

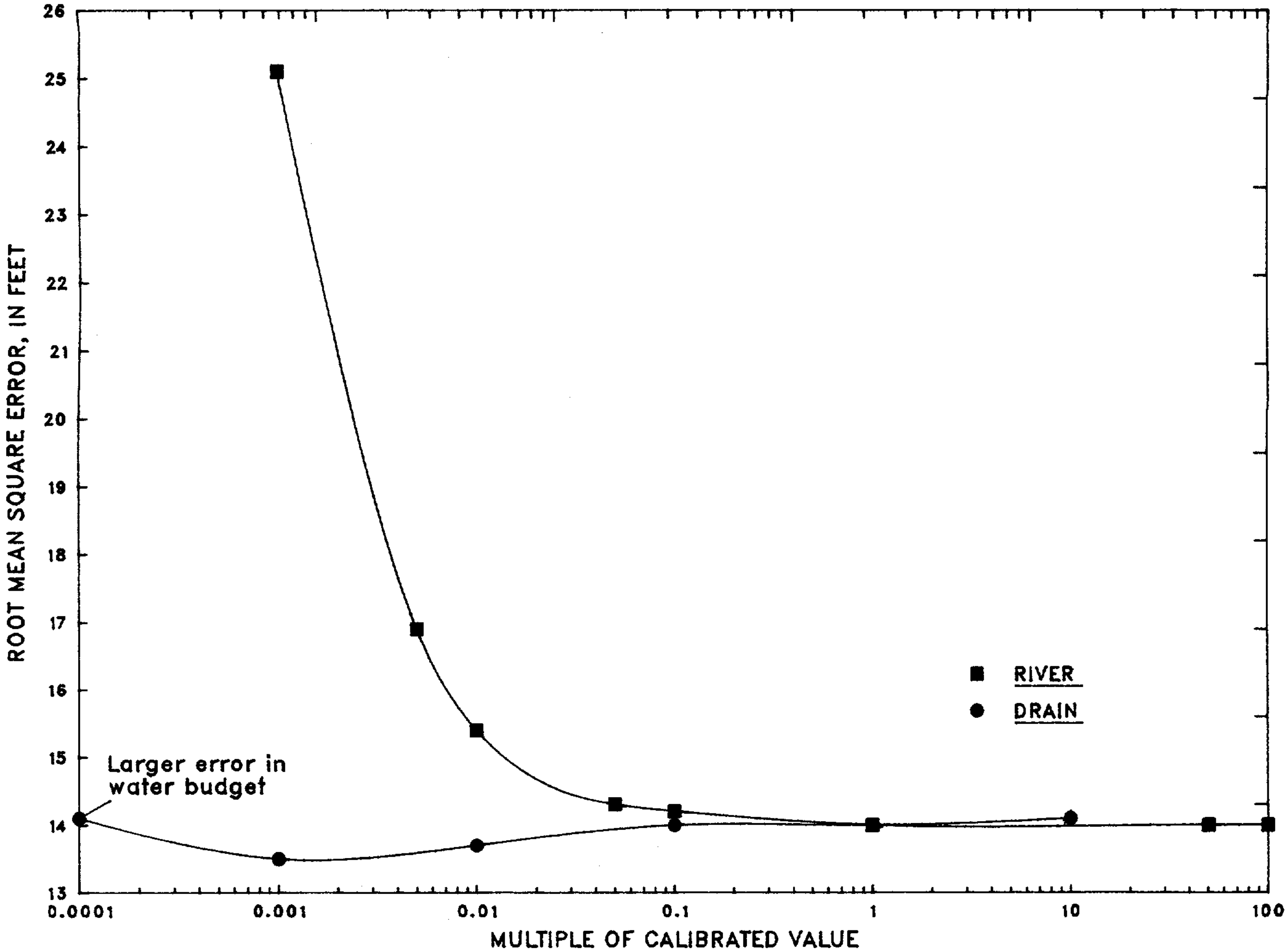


Figure 31.--Sensitivity of the digital flow model to adjustments in drain and river conductance.

matches were improved in a few locations, particularly in areas of steep gradient. These slight improvements in certain areas may indicate that anisotropy is localized and variable, which is probable in a system where secondary permeability features dominate the flow system. However, data are lacking for localized changes in anisotropy, and the model code simulates only uniform anisotropy in a layer.

CONCLUSIONS

Ground water in Bear Creek Valley and its eastern extension, Union Valley, flows primarily from the ridges, which are the primary recharge areas, toward main streams on the valley floor. Potentiometric and geochemical data indicate that ground water in the valley flows across geologic units. There is virtually no difference between water levels or water chemistry in the regolith and shallow bedrock. Both potentiometric and geochemical maps indicate a continuous flow system to a depth of at least 50 feet. Short flow paths along strike (parallel to the valley axis) are controlled by the closely spaced tributaries that are normal to strike or by localized fractures and solution features oriented along strike; however, the principal direction of ground-water flow is from the ridges

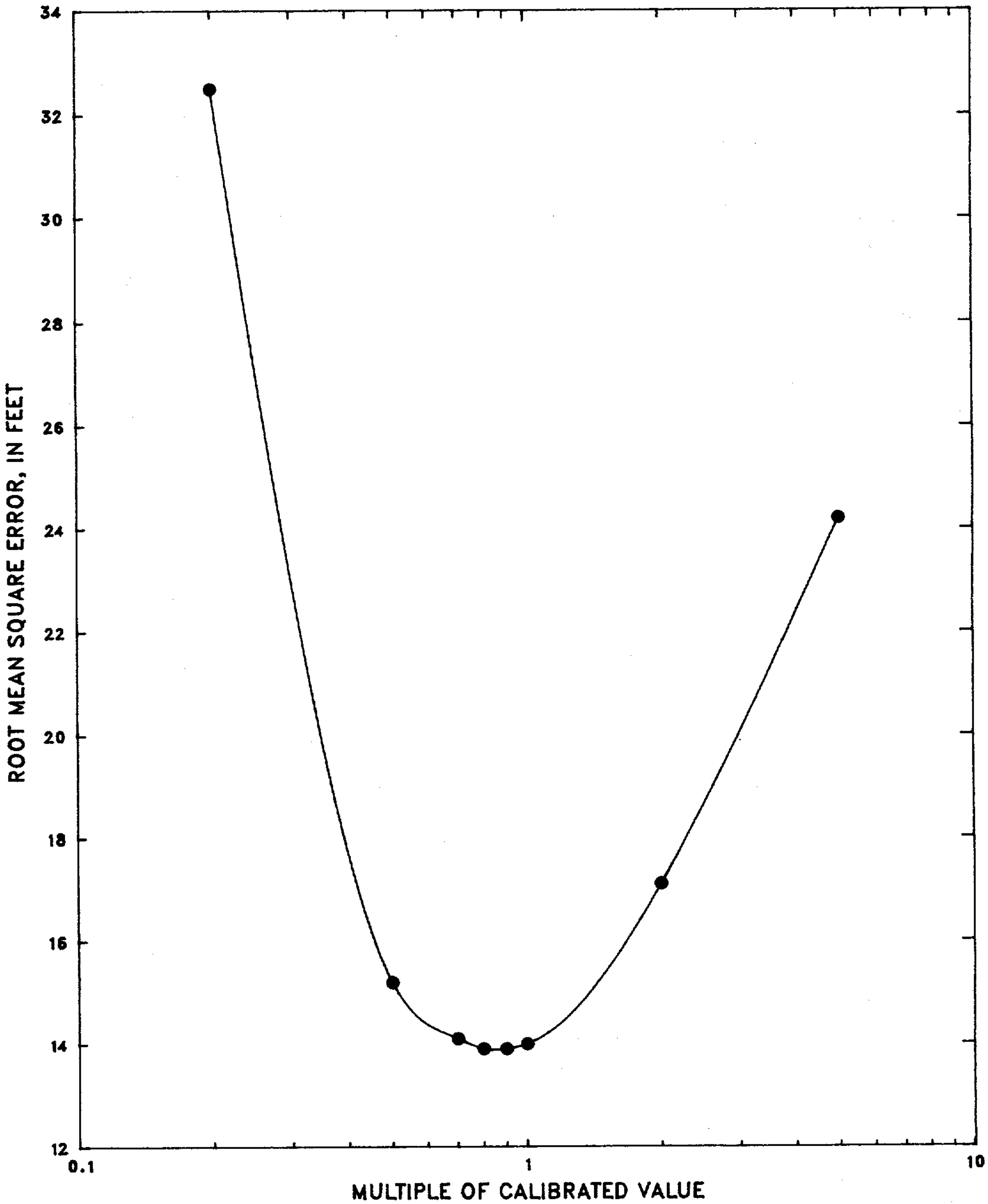


Figure 32.--Sensitivity of the digital flow model to adjustments in horizontal anisotropy.

toward the valley. Preferential flow along strike for any great distance is not indicated, except in the Maynardville Limestone. Flow in the 50- to 500-foot-depth interval is also continuous across geologic units, and a hydraulic and geochemical gradient exists between the shallower and deeper flow.

Ground water discharges in the valley to major streams that flow along the axis of the valley; these streams flow primarily on the Maynardville Limestone, which is fractured and contains solution cavities. The streams function as ground-water drains, although they may temporarily lose water in reaches where fractures are intense or solution cavities are shallow and numerous. This ground-water and surface-water flow in the Maynardville Limestone makes that formation a more likely pathway for potential contaminant transport along strike (down the valley) than other formations.

The limited chemical data available indicate at least two geochemically distinct ground-water zones—one less than 50 feet deep and another between 50 and 500 feet. Water enters the subsurface through permeable zones on ridges and valley slopes and flows toward Bear Creek. Discharge to Bear Creek appears to be principally from the shallow water-bearing zone because the creek water is chemically similar to ground water closer to recharge source. Ground water from the Maynardville Limestone contains a higher percentage of water from the Copper Ridge Dolomite than from the Rome Formation.

Although the chemistry of water from the shallow zone is influenced by naturally occurring water-rock interactions such as dolomite and gypsum or anhydrite dissolution, as indicated by geochemical models, some wells are affected by disposal of acidic, mineral, and organic-process wastes, and by road salting. These activities have caused local dissolved-solids loading of the shallow water-bearing zone, in some instances resulting in concentrations of dissolved solids that exceed 10,000 mg/L. Geochemical data for water in the Rome Formation, Maynardville Limestone, and Copper Ridge Dolomite indicate that some analyses were affected by grout used in well construction and some by road salts.

Water from the deep water-bearing zone (50 to 500 feet) may also discharge to Bear Creek. The chemistry of water from the deeper zone is mostly influenced by natural chemical evolution of ground water, but some contamination in the same areas as in the shallow water-bearing zone is indicated by elevated concentrations of dissolved solids and dissolved calcium.

A four-layer digital model was constructed to simulate steady-state ground-water flow in Bear Creek Valley. The model was calibrated to average water levels, most of which were measured in October 1986, and to ground-water discharge to streams. Areal recharge comprises 97 percent of the inflow to the modeled system; 3 percent of the inflow is leakage from streams. No inflow from outside the valley was assumed, although model results indicate that agreement to measured hydraulic heads could be improved by a source of ground-water influx beneath Pine Ridge. All of the outflow from the modeled system was to streams: 5 percent was through model boundaries at the Clinch River (represented by constant-head nodes), 72 percent through seepage to the main streams in the valleys (represented by river nodes), and 23 percent to tributaries (represented by drain nodes).

Hydraulic conductivity of the formations, determined by statistical and regression analyses of aquifer-test results in early phases of the investigation, were not significantly changed for the three-dimensional model. Average conductivity for the upper 400 feet of strata ranged from 0.0016 ft/d to 0.3 ft/d. Strata below a depth of 400 feet were assumed to have a hydraulic conductivity of 0.000078 ft/d.

Model results and sensitivity analyses indicate that the Maynardville Limestone, because of its (1) position in the flow system, (2) proximity to disposal areas or ground-water flow from disposal areas, and (3) locally high hydraulic conductivity caused by fractures and interconnected solution cavities, is more likely than any other formation to provide a pathway for contaminant transport over long distances. These conditions are particularly prevalent in the area of the East Fork Poplar and Scarboro Creeks drainage divide. The ground-water gradient at depth in this area indicates that contaminants moving with the water could be transported off the eastern end of the Oak Ridge Reservation property. In the Bear Creek watershed, contaminants transported by ground water to the Maynardville Limestone could be discharged to Bear Creek and transported down the valley by the stream or transported by ground water along the strike of the formation through fractures and solution openings. It should be emphasized that the proposed pathways are based on the results of an areal model of the flow system. The interpretation assumes that contaminants (1) move with the ground water and (2) follow the regional flow gradient. The pathways, therefore, are regional in scale and local variations that are potentially very important, are beyond the scope of this study and were not investigated.

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APPENDIX A

Geochemical data from wells in Bear Creek Valley

[Concentrations are dissolved and in milligrams per liter unless otherwise specified. ROE = residue on evaporation at 180 C, uS/cm = microsiemens per centimeter]

Well No.[a]	Date sampled (year/ month/ day)	Well depth (feet)[1]	Temper- ature (°C)	pH	Conduc- tivity (µS/cm)	Dissolved solids (ROE)	Ca	Mg	Na	K	HCO 3	CO 3	Cl	SO 4	NO 3	SiO2
Copper Ridge Dolomite																
GW-165a	860130	280	--	--	--	270	11.8	12.0	14.0	7.9	79.3	--	97.0	1.0	--	0.1
GW-166a	860130	380	--	--	--	120	5.5	7.7	8.0	3.1	56.1	--	32.0	1.0	--	.1
Maynardville Limestone																
1005	850809	8.0	25.9	6.0	--	4,230	219	30.1	118	6.6	--	--	19	--	3,838	2.6
BG-6	850812	47	16	7.3	350	180	61.8	6.6	3.5	1.8	180	--	6.2	5.73	--	2.0
GW-45	860115	3.0	19	8.8	100	130	71	4.9	2.5	3.2	44	--	20.4	--	--	3.2
GW-54	860829	35.2	18	7.0	600	290	86	12	2.0	--	250	--	48	15.0	--	3.3
GW-56	860829	53.2	10	6.5	800	400	44	12	18	--	300	--	150	27.0	--	1.7
GW-57	860516	20.8	16	6.2	163	330	110	27	1.1	1.0	250	--	44	21.0	--	2.1
GW-58	860519	42.2	14	7.6	510	260	78	32	5.3	1.3	170	--	19	36.0	--	3.5
GW-60	860908	47.8	18	6.9	425	290	100	19	11	--	260	--	11.9	20.0	--	3.8
GW-61	860519	19.6	18	7.2	455	280	94	14	5 9	1.7	260	--	12	19.0	--	2.8
GW-62	860515	49.4	17	6.9	580	280	72	27	3 4	.9	260	--	24	19.0	--	4.0
GW-63	860515	27.7	12	6.8	600	330	97	25	10	1.7	280	--	20	35.0	--	3.4
GW-64	860904	50.7	11	6.9	480	350	100	19	6.2	--	240	--	12	85.0	--	5.1
GW-65	860516	29	21	6.7	600	360	100	22	7.6	1.0	250	--	14	86.0	--	2.3
GW-66	860829	52.9	16	6.3	700	470	110	17	7.4	--	350	--	8.3	150.0	--	3.7
GW-67	860829	11.2	13	7.5	360	360	190	13	10	--	130	--	66	16.0	--	5.0
GW-125	861206	502	21	8.9	470	830	1.6	1.6	550	31	240	92	15	16.0	--	.7
GW-167[b]	870423	25	15	6.8	490	300	91	5.1	7.7	2.6	290	--	19	8.0	3.6	9.2
GW-168[b]	870423	104	16	7.3	540	330	82	25	8.7	2.6	370	--	13	7.4	2.4	12
GW-170[b]	870430	104	15	7.8	430	244	45	23	14	2.2	250	--	13	15	3.0	9.8
GW-171[b]	870429	26	14.5	6.9	440	234	55	14	10	3.4	250	--	17	4.2	.4	8.3
GW-172[b]	870430	105	15	7.3	625	373	86	25	14	2.3	360	--	42	1.2	.1	12
GW-212	870420	8.0	15	6.8	320	243	51	27	1.5	.7	220	--	3.4	6.4	.1	11
GW-225	860902	150	14	6.8	800	400	130	29	11	--	420	--	21	27.0	14.2	4.7
GW-226	861216	45	14	6.2	750	360	130	30	9.5	2.9	250	--	26	37.0	--	3.2
GW-227	860114	30	18	7.4	975	440	160	30	16	2.5	280	--	60	30.0	--	3.7
GW-228	860520	80	--	--	--	350	54	28	8.7	5.6	323	--	18	53.0	5.0	7.4
GW-229	860521	40	--	--	--	630	200	27	36	7.2	558	--	86	22.0	--	5.3
GW-230[b]	870429	341	16	7.2	890	536	120	31	20	3.2	440	--	91	2.4	.1	17
GW-236	860905	13	19.6	6.4	310	860	390	46	5.6	--	380	--	210	13.0	--	4.7
GW-237	861218	8.0	14	3.7	27,000	--	--	--	--	--	--	--	--	--	--	--
GW-238[b]	870421	100	15	8.1	430	255	57	22	3.7	3.4	280	--	7.0	16	2.8	9.0

Geochemical data from wells in Bear Creek Valley—Continued

Well No.[a]	Date sampled (year/month/day)	Well depth (feet)[1]	Temperature (°C)	pH	Conductivity (µS/cm)	Dissolved solids (ROE)	Ca	Mg	Na	K	HCO 3	CO 3	Cl	SO4	NO 3	SiO2
Nolichucky Shale																
53-1A	860603	22	25.2	6.1	1,820	6,970	3,400	380.0	110	9.7	286	--	17	--	2,913	20
1003	850811	28	21.9	7.9	17,000	1,560	1,360	173.0	19.2	4.0	--	--	--	--	--	6.5
BG-2	850812	19	20	7.4	310	170	52.4	8.0	6.4	2.4	180	--	5.0	--	--	3.3
BG-3	850809	51	25	5.5	100	50	16	2.1	2.5	1.5	40	--	2.8	--	--	1.1
BG-4	850813	37	16	6.5	800	420	168	12.6	9.3	1.7	360	--	24.9	18.8	--	6.8
BG-5	850811	37	23	11.4	700	220	55.3	--	6.7	15.1	--	140	3.1	--	--	2.9
BG-8	850813	35	20	6.6	200	80	--	3.6	6.5	1.9	114	--	3.1	--	--	4.2
BG-11	860517	35	15.5	7.7	265	150	46	6.2	4.8	1.5	140	--	--	17.0	--	8.8
BG-18	860113	110	11	7.2	750	--	130	17	34	3.2	160	--	140	28.0	--	8.5
GW-3	860515	21.8	17.5	7.6	280	150	47	4.7	7.6	1.4	110	--	33	--	--	4.7
GW-6	860112	37.3	13	6.8	325	220	77	6.1	5.6	2.2	180	--	20	9.8	--	8.1
GW-7	860112	12.3	14	5.7	125	70	19	2.2	5.4	1.6	30	--	26	--	--	6.0
GW-10	861217	7.7	14	6.8	420	--	--	--	--	--	--	--	--	--	--	--
GW-11	860110	40.3	12	7.6	219	--	40	4.9	18	3.3	140	--	14	6.8	--	7.5
GW-12	860828	11.5	18.4	6.7	155	80	13	2.4	5.4	--	80	--	8.6	--	--	9.7
GW-13	860114	8.4	11	7.0	380	310	120	15	4.2	2.1	310	--	8.4	--	--	8.6
GW-14	860110	10	12	6.8	750	920	180	20	9.5	3.4	350	350	170	--	--	11.0
GW-16	860515	13.9	16	6.8	300	170	66	3.8	2.9	1.0	190	--	--	--	--	6.3
GW-18	860114	15.7	13	5.5	25	20	1.9	.8	3.9	2.0	4.0	--	--	7.2	--	5.4
GW-20	860110	55.9	10.5	6.8	920	740	200	17	30	3.3	240	--	360	--	--	8.1
GW-21	861218	11	13	6.65	460	--	--	--	--	--	--	--	--	--	--	--
GW-46	860517	8.1	16	8.4	50	50	17	3.9	4.2	1.3	32	--	4.0	--	--	7.4
GW-47	860115	18.5	16	8.9	460	170	6.2	.98	6.6	.93	230	--	10.5	17.8	--	14.0
GW-68	860109	71.9	13	9.3	255	210	45	12	24.	7.0	88	44	28	--	--	6.6
GW-69	850812	89	11	11.	600	400	8.9	6.3	57.7	3.9	4.0	270	30	20.0	--	3.9
GW-70	860110	125	16	11.4	1,675	950	2.2	--	120	9.8	--	740	27	50.0	--	4.3
GW-71	860515	198.4	15	10.	450	440	1.6	--	400	8.8	--	--	11	10.0	--	5.3
GW-72	860519	87.8	12	7.8	260	220	1.9	--	110	3.0	150	--	14	14.0	--	3.1
GW-73	860515	69.8	23	8.5	550	280	46	7.2	17	3.0	192	88	3.4	12.4	--	6.2
GW-74	860521	180	13	9.6	680	470	.74	--	150	1.0	360	120	--	19.0	--	3.5
GW-75	860519	180	14	7.8	230	330	1.0	--	240	4.0	140	--	5.5	11.0	--	3.2
GW-77	850814	90.3	14	7.8	215	140	33.8	7.8	5.8	2.5	140	--	--	13.0	--	4.9
GW-78	850813	16.1	18	7.9	190	120	41.2	5.9	4.6	1.6	110	--	3.2	7.3	--	2.4
GW-83	860515	24.5	12.5	7.2	210	120	35	2.4	2.7	--	140	--	--	8.6	--	5.2
GW-85	860115	53.8	18	7.1	170	280	170	16	8.7	2.4	140	--	4.2	--	--	7.3
GW-86	860517	24.6	16	6.75	410	190	31	2.2	2.7	.6	200	--	35	9.4	--	7.0
GW-87	860515	9.0	18	8.2	80	130	85	5.7	10	2.2	50	--	2.9	--	--	4.3
GW-93	860828	42	16	9.1	500	200	12	1.7	3.9	--	220	60	--	13.0	--	7.1
GW-94	860518	93.8	15	8.8	440	310	1.4	.74	120	2.1	236	24	4.0	35.8	--	4.0
GW-95	860517	134.8	18	11.8	2,700	250	1.1	--	120	2.2	--	80	14.3	25.8	--	3.5

Geochemical data from wells in Bear Creek Valley—Continued

Well No.[a]	Date sampled (year/month/day)	Well depth (feet)[1]	Temperature (°C)	pH	Conductivity (μS/cm)	Dissolved solids (ROE)	Ca	Mg	Na	K	HCO3	CO 3	Cl	SO 4	NO 3	SiO2
GW-96	860115	44	17	6.65	650	260	20	2.2	4.5	--	400	--	6.1	25.0	--	5.4
GW-97	860828	11.8	12	7.0	350	340	180	4.6	2.0	--	180	--	23	20.0	--	5.8
GW-98	860112	82.4	14	6.5	1,650	360	64	13	16	5.6	210	--	24	130.0	--	7.0
GW-100	860122	10.2	20	6.0	34,000	560	300	24	7.4	2.2	270	--	92	--	--	3.8
GW-101	860529	12.3	17.4	4.0	15,000	5,800	2,400	180	68	5.2	--	--	267	830	2,054	4.4
GW-103	860604	20	18.9	4.1	19,500	9,200	1,500	280	620	53	--	--	224	--	6,457	23.0
GW-104	860604	59.8	17	5.4	10,500	5,000	3,500	590	790	130	--	--	40	--	--	12.0
GW-105	860904	12.1	14	7.2	7,900	1,770	460	50	45	--	--	--	47	11.0	1,158	3.9
GW-106	860904	61.9	14	6.5	1,500	2,000	1,400	180	23	--	490	--	14	350	--	6.6
GW-107	850815	8.5	20.2	7.0	1,780	1,260	392	14.9	18.8	1.5	638	--	7.0	462	42.7	9.4
GW-108	850815	46.7	22.4	6.4	--	30,480	16,600	812	215	28	592	--	60	--	12,463	5.9
GW-109	860519	102.9	17	7.6	360	16,600	13,800	1,900	700	18	230	--	3.4	14.0	--	5.9
GW-116	860116	235	--	10	2,010	30	.96	--	23	3.0	--	--	--	--	--	3.8
GW-117	861220	480	14.5	8.5	280	100	--	--	--	--	180	--	6.9	--	--	--
GW-118	861219	525	15	8.2	310	90	--	--	--	--	170	--	3.6	5.4	--	--
GW-119	861220	460	17.1	9.0	315	12	--	--	--	--	--	--	5.4	6.7	--	--
GW-120	860904	130	10.1	7.1	4,920	990	6.1	1.9	91	--	330	--	76	--	650	5.5
GW-122	861212	92	14.8	7.3	3,130	1,750	820	160	110	25	254	--	140	22.0	343	7.4
GW-123	861223	522	19	10.5	2,080	1,480	1.1	.55	570	40	398	332	230	83.0	25.3	0.4
GW-124	860906	100	12	8.9	550	980	610	66	80	--	290	30	14	33.0	--	6.2
GW-127	860905	19	16.6	9.7	140	370	270	33	15	--	16	16	21	6.0	--	2.5
GW-214[b]	870501	415	16	10.0	3,120	1,820	1.8	.60	730	9.0	370	120	610	--	.1	1.5
GW-239[b]	870501	404	15.5	9.6	2,250	1,310	3.2	.70	520	11.	420	43	430	44.0	.08	3.2
GW-243	861219	45	14	4.4	15,000	--	4,200	--	890	--	--	--	--	--	--	--
GW-244	861220	47	14	5.8	16,000	--	--	--	--	--	--	--	--	--	--	--
GW-245	861220	44	19.9	5.1	13,360	1,770	--	--	--	--	--	--	190	4.5	1,580	--
GW-246	861220	47	19.3	6.9	13,570	2,000	--	--	--	--	--	138	140	20.0	1,761	--
GW-247	861220	47	--	6.2	190	--	--	--	--	--	--	--	--	--	--	--
GW-276	861029	13	13	7.1	170	1,830	1,200	220	350	47.	--	--	--	--	--	8.7
GW-277	861029	67	21	6.2	6,500	3,000	2,500	300	83	35.	200	--	30.8	12.1	--	8.9
Maryville Limestone																
BG-1	850815	28	17	7.7	360	200	63.2	9.1	15.6	1.4	210	--	5.5	--	--	4.9
BG-7	850814	28	16	7.3	280	130	40.7	6.1	6.0	.8	140	--	5.0	--	--	7.5
BG-10	860112	33.3	13	7.6	210	140	44	5.4	4.8	2.0	130	--	--	7.0	--	13.0
BG 15	850812	5.0	24	5.8	65	40	10.5	2.9	3.3	.9	34	--	4.4	--	--	5.5
BG-16	850814	5.0	18	7.0	160	100	17.8	7.3	7.1	1.6	70	--	5.0	24.8	--	6.0
BG-17	860113	100	10	6.8	465	380	120	19	25	3.0	300	--	16	40.0	--	12.0
GW-1	860521	18.1	18.9	7.1	110	40	6.9	1.3	2.0	.6	--	--	5.6	2.7	--	5.7
GW-24	860111	72.6	14	6.4	2,200	1,560	480	55.	56	6.0	250	--	830	--	--	9.8
GW-27	860111	27.3	12	6.2	1,000	690	140	27	75	7.7	240	--	310	--	--	13.0
GW-28	850814	15.2	20	7.5	320	250	62	9.3	5.5	1.7	190	--	63	14.2	--	5.1

Geochemical data from wells in Bear Creek Valley—Continued

Well No.[a]	Date sampled (year/ month/ day)	Well depth (feet)[1]	Temperature (°C)	pH	Conductivity (µS/cm)	Dissolved solids (ROE)	Ca	Mg	Na	K	HCO3	CO3	Cl	SO4	NO3	SiO2
GW-29	860111	15.3	14	6.8	1,450	1,000	280	77	35	5.7	440	--	330	88.0	--	8.6
GW-39	860111	17.9	15	7.3	800	580	180	25	21	3.5	260	--	200	11.0	--	7.1
GW-43	860519	22.8	17	7.5	225	110	13	2.7	5.4	1.3	146	--	4.3	6.4	--	8.6
GW-44	860519	48	18	7.9	305	170	47	3.1	4.0	.5	200	--	6.3	--	--	8.5
GW-82	860908	29.4	13	7.6	150	190	50	5.9	3.5	--	180	--	--	6.4	--	9.5
GW-84	860829	22.8	12	7.5	950	120	58	3.1	5.3	--	100	--	--	--	--	7.5
GW-234	860518	10	20.1	6.0	1,900	160	50	11	7.8	1.8	30	--	63	--	--	8.6
Rogersville Shale																
GW-36	860520	34.9	16.8	7.5	320	90	47	7.8	5.4	2.0	--	--	11	12.0	--	5.4
GW-79	860516	59.9	18	6.5	95	80	32	4.0	3.5	1.5	22	--	--	22.0	--	6.3
GW-80	860516	24.7	11	6.7	240	120	3.9	3.0	7.5	1.5	130	--	13	22.0	--	6.1
GW-115	850809	42	15	7.6	335	100	65	7.8	9.1	1.9	--	--	2.9	12.0	--	4.0
GW-126	860831	105	21	6.5	1,350	720	1.0	--	200	--	250	--	230	160	--	3.6
GW-215[b]	870423	10	14.	7.9	335.	209	56	5.0	11	2.8	200	--	2.6	18	.1	16
Rutledge Limestone																
BG-13	850811	5.0	22	5.7	450	130	9.7	10.3	7.1	2.6	70	--	69	--	--	1.8
BG-14	860115	35	14	6.4	150	90	20	4.4	3.4	1.8	78	--	--	11.0	--	6.6
Pumpkin Valley Shale																
BG-9	850815	28	17	6.8	190	140	31.2	9.0	7.2	2.4	120	--	2.9	13.8	--	13.6
BG-12	850811	33	24	5.9	250	90	4.5	7.7	7.5	1.7	22	--	4.7	48.4	--	7.3
GW-40	860517	25.7	18	6.4	140	100	7.8	7.7	9.9	2.2	34	--	--	39.0	--	12
GW-41	860112	36.6	13	6.0	100	80	5.3	5.0	6.9	2.0	18	--	--	37.0	--	14.0
GW-42	860517	24.9	25.3	6.3	71	40	3.9	3.9	2.3	1.3	22	--	--	8.8	--	7.0
GW-81	860111	13.7	16	7.6	240	160	39	12	10	3.7	150	--	--	7.6	--	16.0
GW-162	860830	92	21	9.0	300	80	12	.7	23	--	--	24	2.5	14.0	--	3.6
GW-242	860518	11	--	--	--	12,350	34	--	9.2	1.8	--	--	430	800	11,063	13.0
Rome Formation																
GW-206[b]	870501	11	15	7.4	525	300	58	25	13	7.6	330	--	17	11	.1	17
GW-207[b]	870424	100	16	7.4	580	359	60	40	10	3.1	350	--	2.6	49	.1	26
GW-208[b]	870424	404	16	7.7	600	360	43	40	10	3.0	290	--	2.0	52	.1	26
GW-209[b]	870422	42	14	5.7	38	29	1.4	.4	3.0	3.2	5.9	--	1.0	.1	.4	19
GW-210[b]	870421	104	16	9.4	310	219	32	27	7.0	4.4	140	14	1.0	33	.4	18
GW-211[b]	870423	404	16.	7.8	1,480	1,310	260	52	19	6.1	160	--	2.6	770	.1	21

[1]Depth to top of open interval.
[a]Maps showing well locations, and location coordinates and alternate well numbers can be found in Haase and others, 1987b.
[b]Data collected and analyzed by U.S. Geological Survey. All other data provided by Martin Marietta Energy Systems, Inc, 1987, and written commun.

APPENDIX B

PHREEQE mass-transfer coefficients for the Rome Formation, Maynardville Limestone, and Copper Ridge Dolomite models

[--- means no data were available]

Mineral phase	Mass transfer coefficients				
	Rome Formation		Maynardville Limestone		Copper Ridge Dolomite
	Step 1	Step 2	Step 1	Step 2	
Recharge CO_2	1.80	(GW-209)	1.80	(GW-209)	1.82
Calcite	.44	-1.78	- .038	0.90	---
Dolomite	1.02	---	1.10	---	.25
CaNa-Exchange	.12	.75	.02	.04	---
Chalcedony	-.08	- .001	- .08	- .001	.235
CO_2	1.50	-2.18	1.00	1.70	1.50
Gypsum	.60	9.00	.05	.10	---
NaCl	---	---	---	.60	---
Na, K, Mg.5 Cl_3	---	---	---	---	3.00
CaMg-Exchange	.45	.75	---	---	---